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Geological Environments of the Eastern Arctic

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Petro-Canada

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Introduction

It is not an easy task to describe the geological environment of an area as large as the eastern Arctic. Besides the vastness of the area, there are many aspects to consider. A glance at the table of contents may surprise the reader with respect to the organization of these aspects. Geomorphology, surficial deposits, coastlines and vegetation have been grouped together for one good reason: they are the things that one sees and feels when present in the Arctic. To help the reader in visualizing the nature of the land, references have been made to appropriate topographic maps. After the tangible comes the inferred knowledge: offshore physiography, tectonic framework, geology, glacial history, and seismicity. It is hoped that this document will convey a little of the beauty of the eastern Arctic.

1 Physiography, Surficial Geology, Coastlines, and Vegetation of the Eastern Arctic

Canada can be subdivided physiographically into two distinct and, in many ways, contrasting units: 1) the Shield, a core of ancient igneous and metamorphic rocks, and 2) the Borderlands, an incomplete rim of stratified younger rocks (Bostock, 1976). The Borderlands surround the Shield on its northern, western, and southern sides, either as lowlands, plains, or plateaus made up of flat-lying strata, or as discontinuous mountain ranges and plateaus made up of deformed rocks. The northeastern side of the Shield is elevated, and is, in many areas, in contact with the sea (Figure 1).

Both the Shield and the Borderlands can be divided physiographically into regions, divisions, and subdivisions. In the Borderlands, the limits of subdivisions are based on changes in topography and geology. The Shield, however, is characterized by a much more uniform terrain and in this case, the lithology, the style of geological structures, and the effects of erosion are used to delimit regions and subdivisions.

Parts of both the Shield and the Borderlands occur in the eastern Arctic. The study area comprises two regions and six divisions (Figure 2). These can be tabulated as follows:

	Regions	Divisions
Shield	Davis Region	- Bylot and Baffin islands coastal lowlands and hills - Davis Highlands - Baffin Upland - Hall Upland
Borderlands	Arctic Lowlands	- Lancaster Plateau - Foxe Plain

1.1 The Shield

The history of the present surface of the Shield began with the development and dissection of a peneplain on Archean rocks in pre-Proterozoic time, at least 2,480 million years ago (Cooke, 1931). This erosional surface appears to have been buried and re-exhumed at least once during the Proterozoic Eon (Bostock, 1976). The Shield was then completely covered by early Paleozoic strata and at least until Middle Devonian time (370 million years ago) supplied no sediment for deposition elsewhere. The erosion of overlying strata appears to have been completed, in the eastern Arctic, by Tertiary time (approximately 5 m.y.a.; Bird, 1959). These events and later the scouring by glaciation, the action of erosive forces (Jenness, 1952), and the effects of glacial deposition gave the Shield its present appearance.

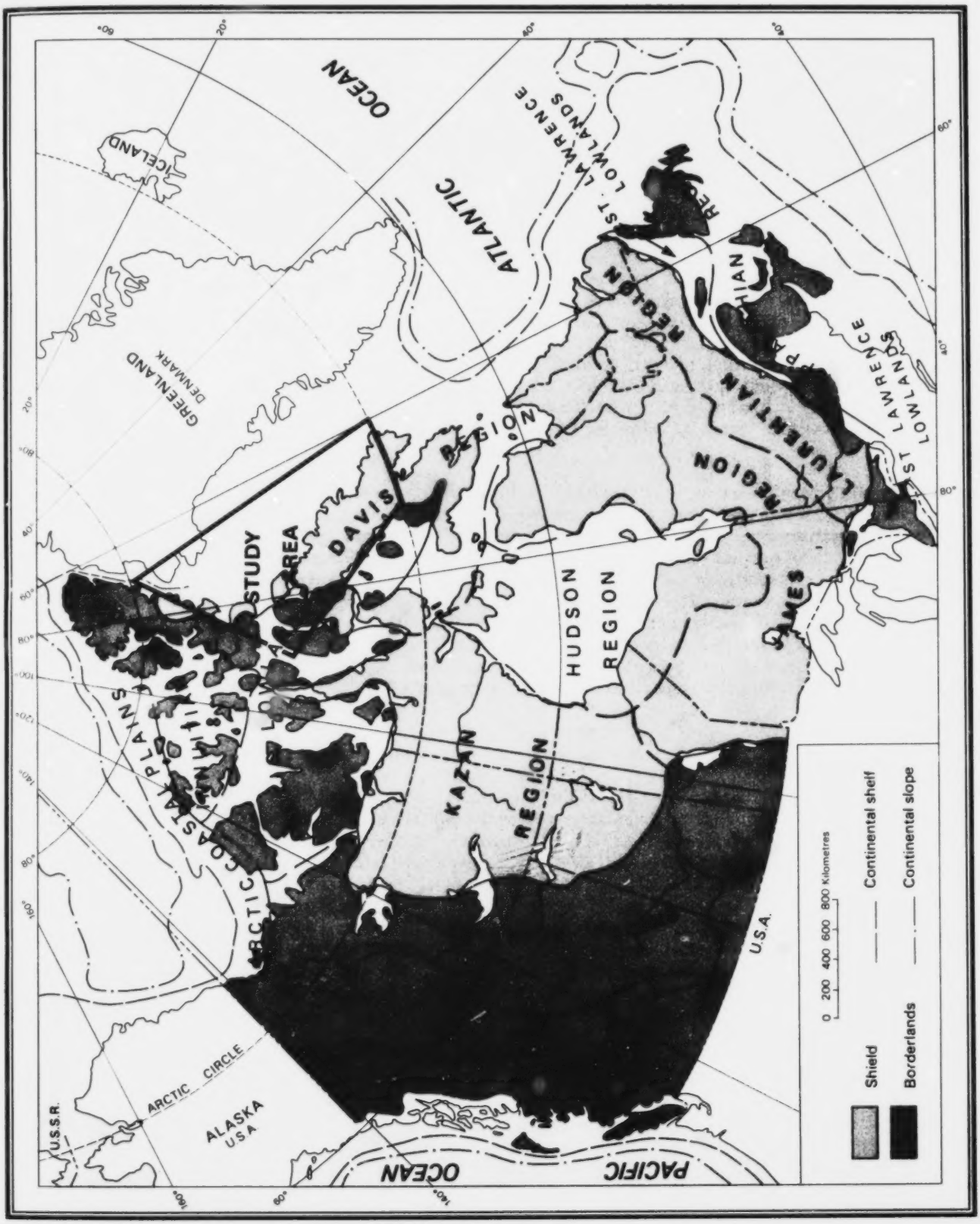


Figure 1 Physiographic Regions of Canada, after Bostock, 1976.

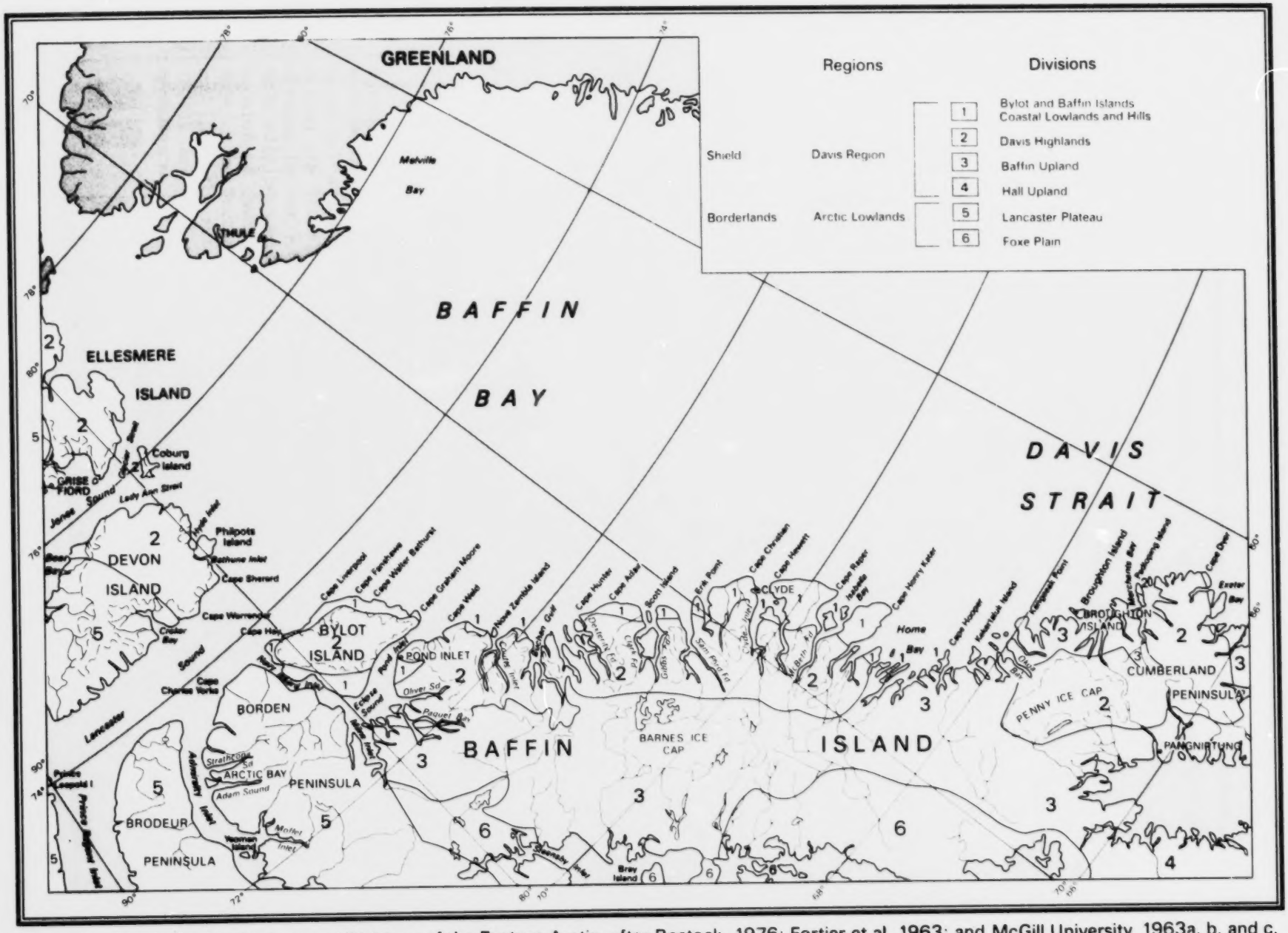


Figure 2 Physiographic Regions and Divisions of the Eastern Arctic, after Bostock, 1976; Fortier et al. 1963; and McGill University, 1963a, b, and c.

Considerable variations exist in the drainage patterns of the Shield, but major features may be as old as pre-Paleozoic (at least 575 m.y.a., Ambrose, 1964). Tributary drainage patterns have, of course, been modified and complicated by the effects of glaciation, but early physiographic studies recognized that the erosional surface "is returning to daylight with precisely the characteristics it possessed when it was buried and preserved ages ago" (Cushing et al., 1910). Furthermore, it seems that the role of glaciation consisted mainly of rejuvenating the existing topography (Baker, 1916, 1939; and Kaszycki and Shilts, 1979), and that it is easy to overestimate the erosive power of glaciers (Lorrain et al., 1981). The main features of the present landscape had developed before the Pleistocene glaciations. Some significant modifications of the preglacial topography by glacial erosion have occurred in the Davis Highlands, but in general, the effects of glaciations and subsequent marine submergence have been restricted to modifying the terrain only in details. The most common modification caused by glaciations is the burial of the erosional surface by glacial debris. These deposits are often responsible for the formation of mosaics of small lakes and ponds on the surface of the Shield. Moraine and drumlin fields often superimpose a strong parallelism on the subsequent development of drainage.

Within the Shield, the Davis Region (Figure 1) extends from Ellesmere Island in the north to Labrador in the south. Relatively narrow at both its northern and southern ends, it is considerably wider in its central part, where it extends from the eastern coast of Baffin Island in the east, to the Paleozoic rocks of Foxe Plain in the west. The general appearance of this region is one of a broad, gently warped erosional surface, with few surficial deposits and a high relief along most of its eastern margin. In the eastern Arctic, the Davis Region comprises four major divisions (Figure 3): the Bylot and Baffin islands coastal lowlands and hills, the Davis Highlands, the Baffin Upland, and the Hall Upland (Bostock, 1976; and McGill University, 1963a, b, and c).

1.1.1 Bylot and Baffin Islands Coastal Lowlands and Hills

The Davis Highlands are flanked by hills and lowlands except on southeastern Ellesmere Island, eastern Devon Island, and the area of Buchan Gulf on eastern Baffin Island. These hills and lowlands are underlain, except on southwestern and northern Bylot Island, by the same igneous and metamorphic rocks that form the adjacent highlands. Because of their relatively steep regional slope, their dissection has been rapid, but some accordant hill tops remain present at different altitudes and suggest successive denudational stages. Glacial erosion has contributed to the dissection of the terrain and is responsible for the presence of many steep-troughed walls. In general, glacial and marine deposits form the surface of the lowlands and obscure the underlying bedrock, except where it protrudes as rounded hills.

Five major subdivisions make up the Bylot and Baffin islands Coastal Lowlands and Hills. These are: the Bylot Island Coastal Lowlands, the Baffin Island Coastal Lowlands, the Bylot Island Coastal Hills, the Baffin Island Coastal Hills, and the coastal area of Home Bay (Figure 4).

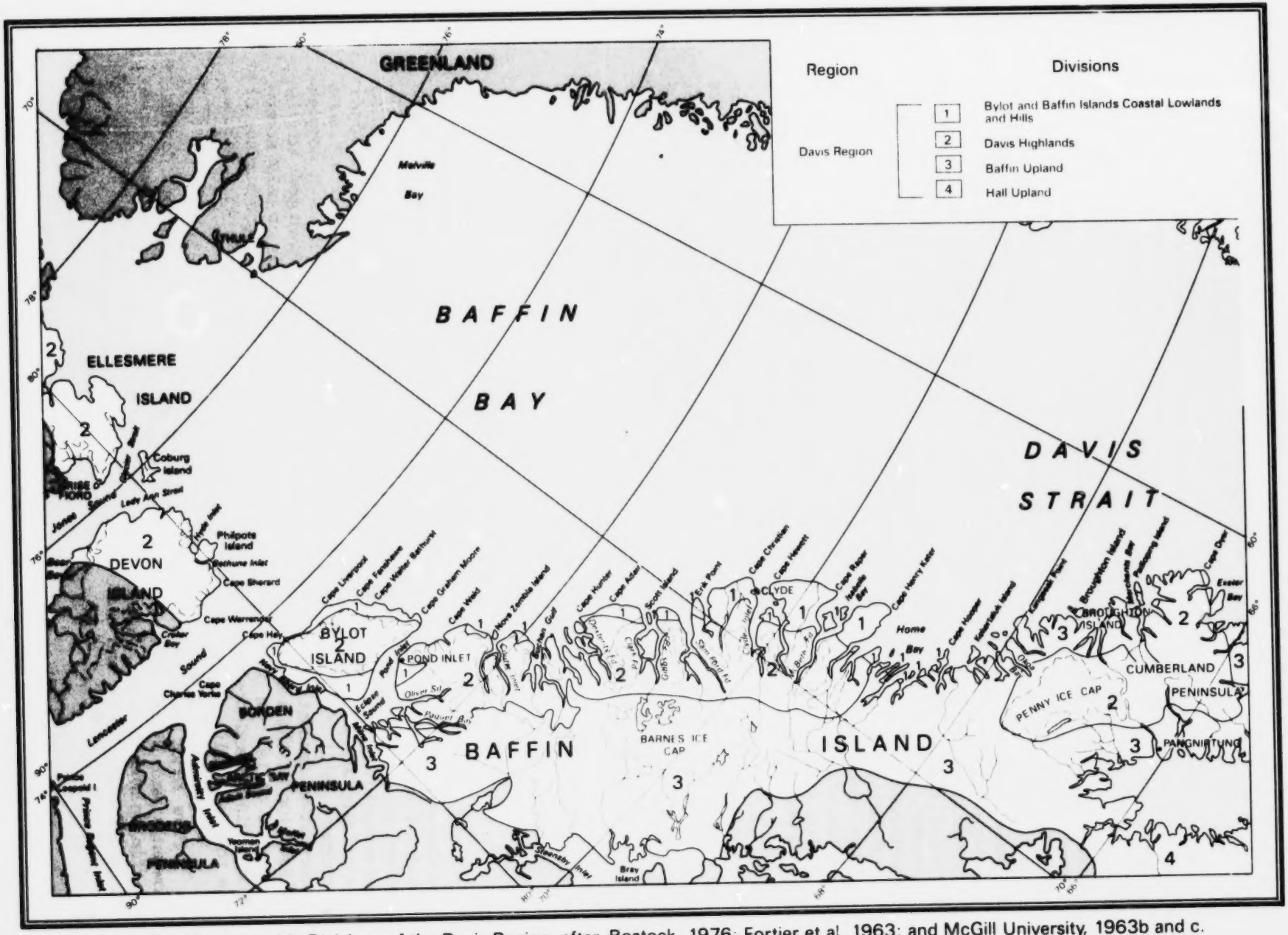


Figure 3 Physiographic Divisions of the Davis Region, after Bostock, 1976; Fortier et al. 1963; and McGill University, 1963b and c.

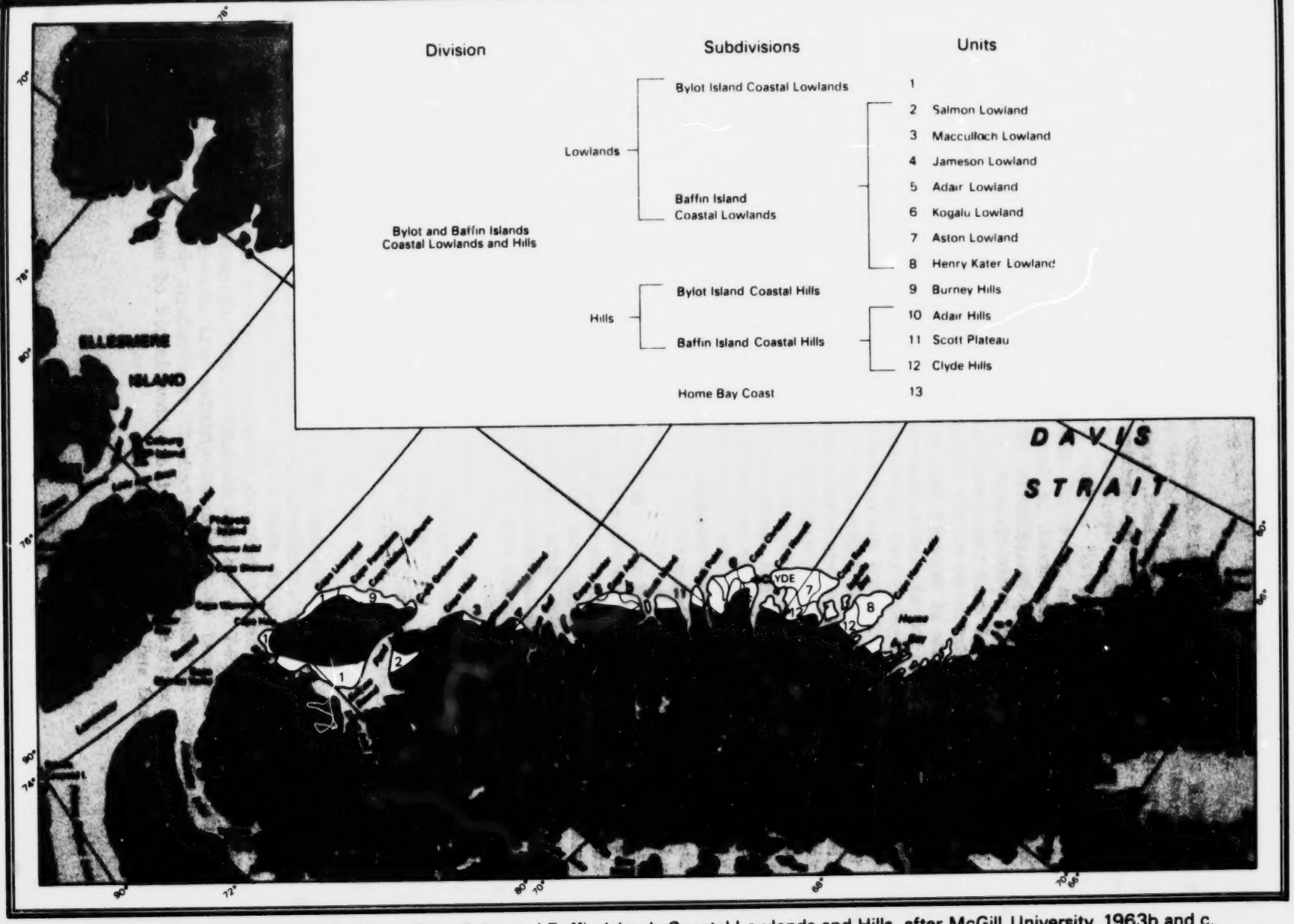


Figure 4 Physiographic Subdivisions of the Bylot and Baffin Islands Coastal Lowlands and Hills, after McGill University, 1963b and c.

1.1.1.1 Bylot Island Coastal Lowlands

The periphery of Bylot Island, from its southwestern corner, around the western and northern coast to Cape Fanshawe, is underlain by sedimentary rocks (Figure 4, Unit 1).

On the southwestern part of the island, a well-developed lowland extends westward from the mouth of Sermilik Glacier, along the foot of the mountains and up to Navy Board Inlet (National Topographic System Maps 48D: Navy Board Inlet, 38C: Bylot Island, and 38B: Pond Inlet). Several broad glacier tongues extend onto the lowland which rises from sea level through a series of coastal bluffs 30 to 60 m high, and slopes gently inland to the base of the mountains at an approximate elevation of 450 m. On the surface of the lowland, braided meltwater streams flow in deep valleys. The streams flowing south to Eclipse Sound cut through older alluvial deposits and form gravel terraces. Further to the northwest, glaciers terminate inland at the foot of the mountains and discharge their meltwater into Navy Board Inlet through two broad valleys that have steep sides and marshy floors. The mouths of these valleys are occupied by deltas, and those of other major streams are occupied by intertidal mud flats. Sand and gravel beaches border the coast. In the southwestern part of Bylot Island, coastal sediments are often backed by marine terraces and raised beaches, sometimes with marshes between successive paleobeaches. Sand dunes occur southwest of Aktineq Glacier, and a few marshes and small lakes occur on the drift covered surface of the lowland.

Inland from Canada Point, a plateau with its summit elevation at approximately 600 m, extends about 10 km inland. This plateau drops steeply to a drift-covered coastal lowland, generally less than one kilometre wide, fringed by continuous gravel and sand beaches. Further to the north along Navy Board Inlet, glaciers from the Byam Martin Mountains extend across the plateau and discharge meltwater into the inlet through a series of troughs, of which the major one is Tay Bay.

On the northwestern part of Bylot Island, the plateau descends to a triangular lowland similar in general appearance to the one present on the southwestern corner of the island. This lowland differs in that it is fringed, in its northern part, by a steep shoreline devoid of sediment. The lowland area occupies approximately 100 km^2 and is bounded on the east by an unnamed glacier that reaches the sea near Cape Hay.

East of this glacier, another plateau flanks the northern side of the Byam Martin Mountains and extends eastward to a glacier that reaches the sea at the western extremity of Maud Bight. This plateau drops to the coast through a series of broad, 150 m high steps that are probably old marine benches that developed on the horizontally stratified bedrock. The benches are partially obscured by a piedmont glacier and are cut by steep-sided valleys. Except in these valleys, bedrock is overlain by drift. Near Cape Hay, the coast consists of steep cliffs either devoid of sediment or fringed by continuous sand and gravel beaches.

Inland from Maud Bight, the surface of the lowland is rolling and cut by steep-sided valleys occupied by braided meltwater streams. Near the coast, these valleys become gorge-like. The coast itself is usually steep and fronted by narrow continuous beaches of gravel and sand, and the eastern corner of the bight is occupied by a large delta. A west-facing blunt promontory, approximately 60 m high, with a few pocket beaches of gravel, forms the eastern limit of Maud Bight.

East of this escarpment to Cape Fanshawe, the general elevation near the coast decreases, relief becomes more subdued, and bedrock outcrops become rare, except along the shoreline near Cape Liverpool. From Cape Liverpool to Cape Fanshawe, gravel and cobble beaches are continuous and often backed by low, eroding bluffs. West of Cape Liverpool, three vertical diabase dykes have formed rocky headlands. Lakes and marshes occur frequently on the drift-covered lowland surface, and near Cape Liverpool, a northwestern deflection of several streams has formed a number of elongated lakes in the backshores of gravel beaches. With the exception of one major glacier tongue that extends for approximately six km on the lowland, all other valley glaciers terminate at the foot of the mountains approximately 10 to 15 km inland.

1.1.1.2 Baffin Island Coastal Lowlands

Eight distinct lowlands occur on the eastern coast of Baffin Island (Figure 4, Units 2 to 8). They range in size from approximately 260 to 1550 km^2, from small coastal areas to large parts of peninsulas. The surface elevation of these lowlands is generally very low and relief rarely exceeds 15 m. Except for the Salmon Lowland, they are underlain by the same igneous and metamorphic rocks that form the adjacent hills, and highlands and are covered by ground moraine and drift material. Low, rounded hills of bedrock sometimes protrude through the overlying unconsolidated sediments, and fluvioglacial, fluvial, and marine deposits occur locally. Because of the low elevations and relief, drainage is often poorly organized, and small lakes, marshes, and swamps occur frequently. A few large rivers flow across the lowlands, often in braided streams. Low, bedrock headlands with or without pocket beaches, raised beaches, marine terraces, low bluffs of till or marine sediments fringed by continuous gravel beaches, and extensive intertidal mud flats, occur along the shorelines of these lowlands.

1.1.1.2.1 Salmon Lowland

A triangular lowland lies south of Eclipse Sound and extends approximately 30 km inland along the Salmon River valley (Figure 4, Unit 2; NTS Map 38B: Pond Inlet). The Salmon River, which drains Utuk Lake, has a broad shallow mouth, and has formed a substantial delta with a wide intertidal zone of sand. The lowland is bounded to the south and northeast by escarpments, the surface expression of faults. The surface of the lowland is covered by drift, with low ridges of concentric end moraines

north of Utuk Lake. Drainage is poorly developed, and lakes and bogs occur frequently. Near the settlement of Mittimatalik (Pond Inlet), a few rounded hills of gneiss protrude through the drift, and in the valley of Salmon River, sandstone outcrops resemble those found across Eclipse Sound on southwestern Bylot Island. The coastlines of this lowland consist of moderately or steeply sloping, unconsolidated backshores fringed by continuous sand and gravel beaches.

1.1.1.2.2 Macculloch Lowland

The Macculloch Lowland occupies part of the eastern extremity of the peninsula that lies between the fiords of Erik Harbour and North Arm, immediately north of Coutts Inlet (Figure 4, Unit 3). This lowland extends approximately from Cape Bowen to Cape Coutts, and its continuity is interrupted by a line of hills that terminate at Cape Macculloch (NTS Map 38A: Nova Zembla Island). A large piedmont glacier lies just west of the area, and meltwater streams originating from this glacier and from other valley glaciers meander across the lowland. These streams are generally braided. The surface of the lowland is composed of drift and is occasionally interrupted by low, rounded projections of bedrock. The very small northern part of the lowland (approximately 5 km^2) southeast of Cape Bowen is similar in general appearance to the larger southern one (approximately 60 km^2). Shorelines bordering this lowland consist of gently sloping, unconsolidated backshores fringed by continuous sand and gravel beaches.

1.1.1.2.3 Jameson Lowland

A small area of lowland occurs between Cape Jameson (Ragged Point) and The Bastions, north of Buchan Gulf (Figure 4, Unit 4). It is approximately 20 km long and two to three km wide, except in its middle section, where it extends approximately 12 km inland along a broad flat-floored valley that carries meltwater from inland ice (NTS Maps 38A: Nova Zembla Island, and 37H: Buchan Gulf). Raised beaches occur frequently near the coast, especially north of The Bastions where they dam an eight km-long lake. Shorelines bordering this lowland consist of continuous beaches of gravel and sand with gently sloping, unconsolidated backshores in the southern half of the area, and gently sloping, low rocky backshores in the northern half of the area.

1.1.1.2.4 Adair Lowland

A discontinuous lowland covers approximately 250 km^2 of the eastern outer coast of Baffin Island, between Dexterity Fiord and Scott Inlet (Figure 4, Unit 5). The lowland occupies the mouths of valleys leading inland, and its continuity is interrupted by hills southeast of Cape Hunter and near Cape Adair (NTS Maps 37H: Buchan Gulf and 27G: Scott Inlet). The gently dipping surface of the lowland is made up of drift and outwash material, and includes many small lakes and marshes. The shorelines present

in this area consist mainly of continuous sand beaches with gently sloping, unconsolidated backshores. Raised beaches occur near the coast throughout the area, and more frequently, northwest of Cape Adair. At and near Cape Adair, the shoreline is steep and rocky, and is sometimes fringed by narrow, continuous sand and gravel beaches.

1.1.1.2.5 Kogalu Lowland

The Kogalu Lowland occupies approximately 650 km^2 on the eastern extremity of the peninsula located between Eglinton Fiord and Clyde Inlet (Figure 4, Unit 6; NTS Map 27F: Clyde Inlet). The surface of the lowland is covered with drift, and slopes gently eastward. Low, rounded hills of bedrock protrude through the drift cover at several locations and form the headland at Cape Eglinton. In general, the lowland is relatively well drained except just north of the settlement of Kangirlugaapik (Clyde) where lakes and marshes do occur. A large river, the Kogalu River crosses the lowland, and many smaller streams also follow an easterly course to the sea. North of the mouth of Kogalu River, the coast is rocky, with pocket beaches of sand and gravel. Southeast of the river's mouth, the coast consists of relatively steep, unconsolidated cliffs, approximately 30 m high, fringed by narrow, continuous sand and boulder beaches. Further to the southeast, and around the northern shore of Clyde Inlet, continuous sand beaches are backed by gently sloping, and moderately sloping unconsolidated backshores.

1.1.1.2.6 Aston Lowland

This area (Figure 4, Unit 7) forms the longest (80 km along the coast) and most extensive (approximately 1,550 km^2) lowland on the eastern coast of Baffin Island. It occupies the eastern extremity of the peninsula lying between Inugsuin Fiord, Isabella Bay and McBeth Fiord, and extends from Halliday Point to Cape Raper (NTS Maps 27F: Clyde Inlet, 27E: Cape Hewett, 27D: Cape Henry Kater, and 27C: McBeth Fiord). The drift-covered surface of the lowland slopes gently eastward from an elevation of about 60 m and includes low hills, themselves mostly drift-covered. In its central part, a series of broad, marshy terraces separated by low bluffs start at a point located a few kilometres west of the coast and extend as far as 12 km inland. Numerous rivers cross the lowland, and two major ones reach the sea north and south of Cape Aston. In general, drainage improves inland but a few large lakes occur near the western margin of the lowland. Low, unconsolidated or rocky bluffs, fringed by continuous gravel beaches, form the northern and southern parts of the coast. Cape Raper is a rocky headland with a few pocket beaches of gravel. The central part of the coast, from south of Cape Hewett to south of Cape Aston, is exceptionally shallow; the shoreline consists of continuous sand beaches with wide intertidal zones, and much of the coast is fringed by a sand bar with a shallow lagoon behind it. The sand bar is breached by the sea in a few places and many streams discharge into the lagoon. Near Cape Aston, these intertidal and shallow zones attain a width of approximately three km.

1.1.1.2.7 Henry Kater Lowland

Henry Kater Peninsula, the northern limit of Home Bay, has at its eastern extremity over 1,000 km^2 of lowland (Figure 4, Unit 8; NTS Map 27D: Cape Henry Kater). The drift surface of the lowland slopes down from an approximate elevation of 100 m in the west to 15 or 20 m-high bluffs at the sea. In the western and southern parts of the lowland, low hills of gneiss protrude through the drift cover. In general, drainage is poorly developed and lakes and marshes are abundant. The southern part of the lowland consists of an uneven sand terrain. Raised beaches occur at the head of bays on the southern part of the coast and northwest of Cape Henry Kater. Most of the shoreline consists of unconsolidated bluffs, 15 to 20 m high, fringed by continuous gravel and sand beaches.

1.1.1.3 Bylot Island Coastal Hills

The eastern coast of Bylot Island, between Cape Fanshawe and Button Point, is occupied by a 90 km-long and 16 km-wide belt of hills called the Burney Hills (Figure 4, Unit 9; NTS Maps 38C: Bylot Island and 38B: Pond Inlet). Along this belt, the Archean and Aphebian metasedimentary, metavolcanic, and igneous rocks of the Byam Martin Mountains reach the coast, and except near Cape Fanshawe and Button Point, have been eroded and dissected into rounded hills. These hills are separated by wide valleys (up to 10 km wide) floored with outwash material. The meltwater braided streams flowing in these major valleys terminate in extensive deltas, lagoons, and systems of barrier beaches. Along the coast, steep rocky backshores, either devoid of sediment or with some pocket beaches of gravel and boulder often border areas where hills reach the sea. These alternate with extensive continuous or barrier beaches of gravel and sand. A few small ice caps occur in the northern and southern parts of the area, south of Cape Fanshawe, east of Cape Burney, and northwest of Cape Graham Moore and Button Point.

1.1.1.4 Baffin Island Coastal Hills

Following the events that formed the highest concordant surfaces on Baffin Island, a series of erosional episodes created a number of broad steps on the seaward slope of the Baffin Highlands. This transition zone between the highlands and lowlands forms a distinct physiographic subdivision: the Baffin Coastal Hills. These hills have been dissected by fluvial and glacial processes and are presently ice-covered in many locations. Gneisses overlain by a thin veneer of bouldery drift underlies most of the hill area. Elevations range from 300 to 400 m and seldom exceed 600 m. Except for the presence of very large erratics and some steep-walled glacial troughs, there is little evidence of glacial events. The general topography of the hills, with their numerous U-shaped valleys and rounded interfluves, is typical of fluvial erosion. Several broad valleys, floored with ground moraine, open eastward to the coastal lowlands. Although some

of these contain small lakes, drainage is, in general, better organized than on the adjacent lowlands. The troughs of glacial origin are characterized by flatter floors and corresponding poorer drainage, lateral or terminal moraines, and outwash material.

1.1.1.4.1 Adair Hills

North of Scott Inlet, an area of hills three to 16 km wide borders the Adair Lowland from southwest of Dexterity Fiord to Clark Fiord (Figure 4, Unit 10; NTS Maps 37H: Buchan Gulf and 27G: Scott Inlet). The hills are composed mainly of horizontally banded gneisses partially covered with a bouldery drift. Bedrock has been dissected by fluvial erosion and drainage has a dendritic character. Local relief varies between 60 and 90 m, and the accordant summits of the rounded interfluves rise to an elevation of 450 m. Two large glacial troughs cut through the area of hills. The northern one trends northeast and contains a 15 km-long lake. Northwest of Scott Island, the second major trough trends roughly west and appears to be structurally controlled. Both troughs cross the peninsula and lead westward to Dexterity Fiord. The northern shore of Clark Fiord consists of sections of steep cliffs devoid of sediment, gently sloping, unconsolidated backshores with continuous gravel and boulder beaches, and steep cliffs with colluvial debris and gravel and boulder beaches. The southeastern shore of Dexterity Fiord comprises sections of gently sloping, unconsolidated backshores with continuous sand, gravel, and boulder beaches; moderately sloping and steep, rocky backshores with colluvial debris, and continuous boulder beaches.

1.1.1.4.2 Scott Plateau

The seaward extremities of the peninsulas lying between Scott Inlet, Sam Ford Fiord, and Eglinton Fiord are characterized by a plateau topography (Figure 4, Unit 11; NTS Maps 27G: Scott Inlet and 27F: Clyde Inlet). The plateau surface is gently undulating and slopes eastward from an approximate maximum elevation of 300 to 450 m at its western end. Near the sea, the plateau terminates abruptly, generally at an elevation of 150 to 300 m. The surface of the plateau is covered by a bouldery drift. A broad glacial valley, the Stewart Valley, crosses the peninsula between Walker Arm and Refuge Harbour on Gibbs Fiord. Most of its length is occupied by a long narrow lake dammed at two locations by terminal moraines bordering glacier fronts. The shoreline of Gibbs Fiord consists of steep rock cliffs either devoid of sediments or fringed by colluvial debris and narrow boulder beaches. Similar shorelines border Walker Arm and Sam Ford Fiord. In Eglinton Fiord, the shores consist of moderately and steeply sloping, unconsolidated backshores often fringed by continuous gravel and boulder beaches. The southeastern shore of Eglinton Fiord has a rocky backshore fringed by a continuous beach of sand, gravel or boulder.

1.1.1.4.3 Clyde Hills

Parts of the peninsulas lying between Eglinton Fiord and Home Bay are occupied by areas of rounded hills separated by broad valleys (Figure 4, Unit 12). Between Eglinton Fiord and Clyde Inlet, the area of hills is irregular in shape, and is separated into a northern and a different southern component by the northeasterly trending Ayr Lake and Kogalu River (NTS Map 27F: Clyde Inlet). North of the Kogalu River, the area is typical of the hill subregion with rounded hills and broad valleys. South of the Kogalu River, the Kogalu Lowland extends southwestward along the river valley, and an area of alpine topography extends the highland region to the east, north of Clyde Inlet. Between the alpine area and Clyde Inlet, a plateau surface reaches an elevation of 450 m and is cut by several narrow valleys. The southern shore of Eglinton Fiord consists mainly of steep, moderately sloping, and gently sloping, rocky backshores with colluvial debris and continuous gravel, boulder, or sand beaches. Unconsolidated backshores with continuous gravel and boulder beaches occur near the entrance of Eglinton Fiord.

The northern shore of Clyde Inlet consists of sections of steep, rocky backshores with colluvial debris fringed by narrow, continuous boulder beaches. These alternate with sections of moderately sloping, unconsolidated backshores fringed by continuous gravel and sand beaches. This last type of shoreline occurs most frequently in front of the relatively narrow valleys that trend northwesterly into the plateau area.

The eastern extremity of the peninsula that lies between Clyde Inlet and Inugsuin Fiord is occupied by a plateau lying at an approximate elevation of 450 m. The shorelines of this peninsula consist of sections of steep rocky cliffs devoid of sediments, moderately sloping, rocky backshores with colluvial debris and boulder and gravel beaches, and gently sloping, unconsolidated backshores with continuous gravel and sand beaches. An unusual steep, rocky, elongated headland projects into the south side of Clyde Inlet, and shelters a rectangular bay with gently sloping, unconsolidated backshores and continuous sand and gravel beaches at its head.

Between Inugsuin and McBeth fiords, the hills lie inland of the Aston Lowland (NTS Map 27C: McBeth Fiord). In this area, the hills are traversed by streams draining north and south, and by a major eastward flowing river. The southern shore of Inugsuin Fiord consists, in the hill area, of steep or moderately sloping, rocky backshores, often fringed by continuous gravel beaches. These are often interrupted by bays that include moderately or gently sloping, unconsolidated backshores and continuous gravel or sand beaches. Except for sections of gently sloping, unconsolidated backshores fringed by continuous gravel or sand beaches in front of valleys opening to the sea, all of the north shore of McBeth Fiord consists, in this area, of steep, rocky backshores, either without sediment or with colluvial debris and continuous boulder beaches.

Aulitivik and Aulitiving islands in Isabella Bay have a plateau surface at an approximate elevation of 300 m (NTS Maps 27C: McBeth Fiord and 27D: Cape Henry Kater). The surface of the plateau is covered by a thin layer of bouldery drift and slopes steeply in the coastal zones. Most of the shorelines of these two islands consist of steep, rocky backshores devoid of sediment, moderately sloping, rocky backshores with colluvium and continuous boulder and gravel beaches, and moderately or gently sloping backshores with continuous gravel and sand beaches.

On Henry Kater Peninsula (NTS Map 27D: Cape Henry Kater), relatively short rivers cross the hills and reach the sea through broad valleys that often terminate in moderately or gently sloping, unconsolidated backshores with continuous gravel and sand beaches. Between these valleys, the coastline consists almost entirely of steep, rocky cliffs, either without sediment or with colluvial debris and continuous boulder or gravel beaches.

1.1.1.5 Home Bay Coast

The eastern and northern extremities of peninsulas projecting into Home Bay and the offshore islands of this bay form a distinct area with many characteristics of a submergent coastline (Figure 4, Unit 13; NTS Maps 27B: Ekalugad Fiord and 27A: Home Bay). Outcrops of igneous and metamorphic rocks underly most of the area and are locally overlain by a thin veneer of bouldery drift. The topography is very irregular, with numerous joints, bassins with lakes, and some cirques. Many summits show some degree of accordance at an approximate elevation of 250 to 360 m. Shorelines are generally steep and rocky, with little or no sediment. Colluvial talus fringed by boulder beaches, narrow pocket beaches of gravel, and narrow continuous beaches of boulder or gravel, occur locally.

1.1.2 Davis Highlands

The Davis Highlands form an elevated belt of deeply dissected igneous and metamorphic rocks that extends along the eastern part of Ellesmere, Devon, and Bylot islands, and borders the coastal lowlands and hills on Baffin Island (Figure 3, Division 2). In its northern part, this physiographic division is discontinuous, and on Baffin Island, it can be described in terms of four distinct subdivisions.

1.1.2.1 Southeastern Ellesmere Island

At the southeastern extremity of Ellesmere Island, the Davis Highlands are largely covered by ice, and their extent coincides with outcrops of Precambrian igneous and metamorphic rocks (Frisch et al. 1978; Figure 5, Unit 1; NTS Maps 39F and 39E: Ekblaw Glacier, 39C and 39D: Talbot Inlet, 49A: Craig Harbour, 39B: Clarence Head, and 48H and 38G: Lady Ann Strait). The highlands are approximately 50 to 100 km wide and extend along

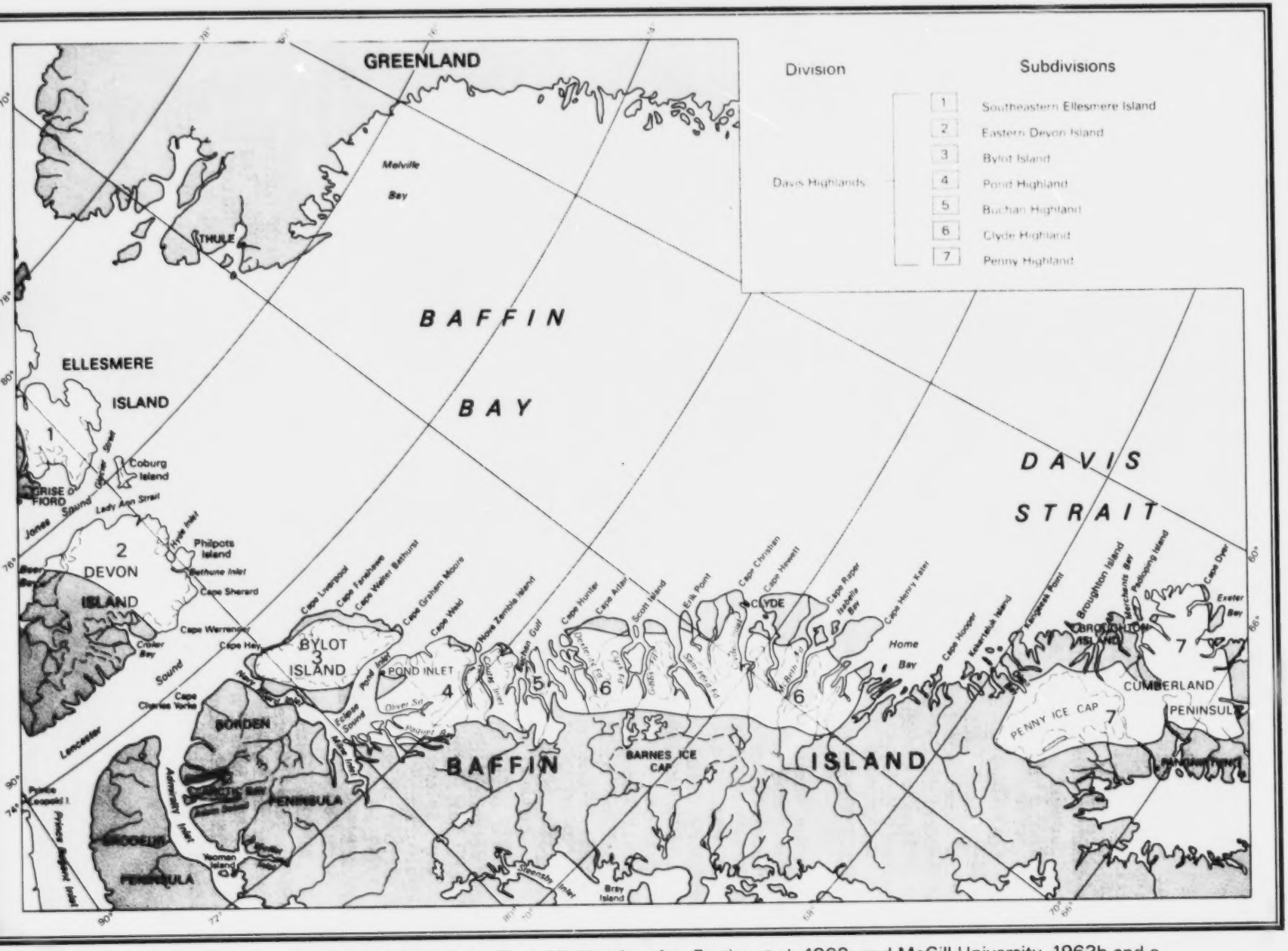


Figure 5 Physiographic Subdivisions of the Davis Highlands, after Fortier et al. 1963; and McGill University, 1963b and c.

the coast between Buchanan Bay and Jones Sound. In the southern part of Ellesmere Island, the highlands extend west to the eastern branch of Starnes Fiord, and a narrow strip continues west along the coast as far as the settlement of Ausuittu (Grise Fiord). To the north of Starnes Fiord, the western limit of the highlands follows a line convex to the west that crosses Makinson Inlet east of the forks, and reaches Buchanan Bay along Beitstad and Hayes fiords (NTS Map 39G: Sawyer Bay; Roots, 1963).

Most of the highlands are covered by an ice cap that reaches an altitude of 1,980 m southwest of Buchanan Bay. The ice cap is drained by glaciers that occupy all main valleys and form an interconnecting network that confines rock outcrops, except along the southeastern coast, to nunataks, valley walls, and coastal bluffs. The rugged topography of the area is expressed through the overlying ice cover at several locations. Irregularities in the unbroken ice cover suggest peaks, ridges, valleys, cirques, branching sharp ridges, and dendritic valleys that could not all have developed under ice. It appears that the ice cap lies upon an area that has been cut by stream erosion and alpine glaciation and may have altered the underlying terrain only slightly. Along the coast, sharp bluffs alternate with glacier fronts, many of which actively discharge icebergs into Baffin Bay. There is very little low, ice-free land.

The highlands are also present on Coburg Island at the eastern entrance of Jones Sound, between Ellesmere and Devon islands. From the air, the island is very conspicuous with high black cliffs and an ice cap that is pierced by many peaks, the highest of which reaches an altitude of 830 m (Dunbar and Greenaway, 1956). Many glaciers flow down from the ice cap along northerly and northwesterly trending valleys, especially on the western and northern sides of the island, where most of the coastline is made up of rounded glacier fronts separated by high, vertical rock cliffs rising abruptly from the sea (Stevenson, 1978). Southeast of the Marina Peninsula on Coburg Island lies a very small, sharp, and high pinnacle of rocks called Princess Charlotte Monument.

1.1.2.2 Eastern Devon Island

On Devon Island, the physiography is linked very closely with the underlying rock type and structures, and there is a direct correlation between physiographic divisions and the tectonic units of the island. The highlands occupy approximately the eastern quarter of the island, (Figure 5, Subdivision 2). The area is largely covered by ice and, from a distance, resembles a steep-sided, flat-topped dome, most of which is lying at an elevation of more than 900 m (NTS Maps 48H and 38G: Lady Ann Strait, and 48E and 38F: Dundas Harbour). The central part of the ice cap reaches an altitude of 1890 m. The boundary between the highlands and the adjoining plateau on the west cannot be defined precisely, as most of it lies under the ice. The highland area can, however, be considered to include all of Devon Island east of a line, convex to the east, that joins Cape Skogn on the northern coast, to Dundas Harbour on the southern coast. It is probable

that this area is completely underlain by Precambrian igneous and metamorphic rocks and is part of the "Baffin-Ellesmere Mountains" of crystalline rocks (Fortier, 1957). The western boundary, where exposed, corresponds to a transition between the cliffed and gullied igneous and metamorphic rocks of the east, and the horizontally stratified rocks on the west. From surface profiles of the ice cap, its uniform drainage patterns, and the elevation and distribution of rock outcrops, it is thought that the ice cover is relatively thin, perhaps 300 to 600 m thick. It is also thought that, contrary to the corresponding highland on southeastern Ellesmere Island, the underlying bedrock surface is smooth, probably dome-shaped, and little dissected. Glaciers draining the ice cap are, in general, short and steep and most discharge into the sea. Many glaciers are confined by steep rock walls or extensive lateral moraines, and a few appear to have actively eroded their substrate. Most of the other glaciers, and the ice cap itself, appear to be presently very passive and to act more as a protective cover than an erosion agent on the underlying bedrock.

At the northwestern corner of the highlands, the terrain is transitional from the plateau to the west, with ridges, hills, small peaks, and gullied bluffs separated by glacier valleys. East of this zone, the coastal bluffs remain steep and rugged, and alternate with glacier fronts as far as Belcher Point. Inland of these bluffs lies the Trueter Mountains, an area of small, glacier-capped mesas, scarps, and angular nunataks. Southeast of Belcher Point, the coast is less steep, has few beaches, and several wide glacier fronts reach the sea. Philpots Island, at the southeastern corner of Devon Island, is anomalous in the context of the highlands as it consists entirely of a low (less than 100 m), rocky, ice-scoured peninsula, the only extensive lowland in the area. Unusual too are the beaches of Philpots Island. They consist mostly of pocket and barrier beaches of boulder and gravel backed by gently and moderately sloping, rocky backshores, and by lagoons. South and west of Philpots Island, the land rises more gently, and a 12 km-wide lobe of the ice cap reaches the sea south of Bethune Inlet. Between Bethune Inlet and Cape Sherard, a gently sloping, rubbly foreland five to eight km wide separates the Cunningham Mountains from a low, rocky coast with pocket beaches of boulder and gravel. Along the southern coast, the Cunningham Mountains form a range of interconnected sharp ridges with gullied flanks and V-shaped valleys associated with extensive lateral and medial moraines and outwash deposits. Four major valley glaciers descend from the ice cap, and their fronts interrupt the continuity of gravel and sand beaches that occur frequently in this area. A large gravel spit is being constructed immediately west of the front of Phoenix Head Glacier and continuous beaches of well-sorted sand occur between Cunningham West and Phoenix Head glaciers. From Cape Warrender to Dundas Harbour, the coast is characterized by steep cliffs that rise above Lancaster Sound to elevations exceeding 900 m less than two km inland. Narrow, continuous pocket beaches of gravel alternate with steep, rocky headlands, and the only occurrences of gently sloping, unconsolidated backshores are restricted to sections of coastlines located immediately in front of glacial valleys.

1.1.2.3 Bylot Island

On Bylot Island, the central Byam Martin Mountains link the highlands of southeastern Ellesmere Island and eastern Devon Island with the one present on Baffin Island (Figure 5, Subdivision 3). The highlands cover approximately 8,000 km^2 or 67 per cent of Bylot Island, and are characterized by snow fields, numerous valley glaciers, high elevations (next to Cumberland Peninsula, the highest parts of the Davis Highlands occur on Bylot Island, where peaks reach altitudes of 1,500 to 1,800 m) and steep relief (NTS Maps 48D: Navy Board Inlet, 38C: Bylot Island, and 38B: Pond Inlet). Most of the highlands are ice-covered and the bedrock forming the northwesterly trending mountains outcrop as deeply eroded, sharp nunataks, arêtes, cols, tors, and cirques. The mountains reach the coast along Pond Inlet and elsewhere, they are bordered by lowlands and hills (Figure 4). Numerous glaciers flow outward from the backbone of the Byam Martin Mountains, and some of the larger ones are excellent examples of valley glaciers with moraines and sandurs (Jackson and Davidson, 1975). Three major glaciers calve into the sea, two on the northern coast near Cape Hay, and one, the Sermilik Glacier, in the centre of the southern coast. Other glaciers terminate in valley tongues or expand into lobes that sometimes coalesce onto the bordering lowlands. Lateral, terminal, and medial moraines are often conspicuous in the lower parts of these glaciers, and have material derived from the inland metamorphic terrain (DiLabio and Shilts, 1978). Most of these glaciers appear to have remained stationary or to have retreated slightly during the recent past. Streams are usually short and glacier-fed, and are often braided throughout much of their course.

1.1.2.4 Eastern Baffin Island

On eastern Baffin Island, the Davis Highlands occupy a 90 km-wide belt of deeply dissected igneous and metamorphic rocks that extend southward from Eclipse Sound and Pond Inlet to 60° N, the latitude of Home Bay. At that point, it becomes discontinuous and it reappears at the latitude of the Penny Ice Cap where it covers approximately 3,000 km^2 of the southeastern part of the island (Figure 5, Subdivisions 4 to 7). The highlands are characterized by deeply penetrating, vertically-walled fiords, ice-covered plateaus, mountains separated by numerous valley glaciers, a steep relief, and by very spectacular scenery. Many peaks reach elevations of 1,500 to 1,800 m, and in general the highlands rise relatively abruptly from the lower land masses both to the west and to the east. Within the highlands area, high-level benches along the fiords indicate late stages in the denudation of the land, if one assumes that the valleys were occupied by rivers leading to a sea level higher than the present one. It is probable that, during the peak of the last glaciation, the areas of highlands were completely ice-covered, except perhaps for a few high peaks. This ice cover must have been drained by numerous valley glaciers and as sea level then was lower than the present one, these glaciers, must have been able to overdeepen their valleys. During the retreat of the last glaciation, the land was depressed by the weight of the remaining, ice and shorelines were

located higher on land than where they are today. Subsequent stages in the isostatic uplift of the land formed river terraces and deltaic accumulations at the head of fiords. On Baffin Island, the Davis Highlands can be described in terms of four distinct subdivisions: the Pond, Buchan, Clyde and Penny highlands.

1.1.2.4.1 Pond Highland

The Pond Highland is a massive plateau of more than 15,000 km^2, cut into gneisses and other metamorphic rocks, with approximate elevations of 1,500 m in its northern and eastern parts and 600 m in its western part (Figure 5, Subdivision 4). The highland area lies north of a line joining Buchan Gulf, Quernbiter Fiord (eastern Baffin Island) to Paquet Bay (southern Eclipse Sound) together with two small areas: one between Tay Sound and Paquet Bay, and the other west of Tay Sound (NTS Maps 38B: Pond Inlet, 38A: Nova Zembla Island, 37G: Icebound Lakes, and 37H: Buchan Gulf).

Within this overall area, a small (approximately 40 km^2) but distinct plateau lies between the Salmon Lowland, Oliver Sound and Eclipse Sound. Its elevation decreases westward from about 550 m near Utuk Lake, and most of the drainage is carried westward by rivers flowing in shallow valleys to Eclipse Sound. To the north, the plateau is bounded by a northwesterly trending fault-line scarp that reaches Eclipse Sound between Tunuiaqtalik Point and Mittimatalik (Pond Inlet). To the west, sedimentary rocks and coal (of possible Jurassic age) are known to occur in the valley of the Patricia River. This suggests that the plateau surface may correspond to, or lie close to, the unconformity that once separated the rocks of the Shield from overlying sedimentary strata. In its higher reaches, the plateau is covered by a veneer of drift. To the east near the coast, the slope steepens and bedrock headlands become frequent at the shoreline.

In Oliver Sound, sections of steep, rocky backshores with some narrow boulder and gravel beaches alternate with short ones of moderately sloping unconsolidated backshores with continuous beaches of gravel. Around Eqeperiaqtalik and Tunuiaqtalik points, gently sloping, rocky backshores and headlands with pocket beaches of gravel, and moderately sloping unconsolidated backshores with continuous sand or gravel beaches, form the shorelines.

Elsewhere than on this plateau, the highland has elevations that reach 1,525 m, has a much steeper relief and is generally more dissected. The surface consists of bedrock, sometimes mantled with boulders, or in the highest parts, ice; especially south of Oliver Sound where an extensive ice cap is located. Numerous outlet glaciers are present, and many reach the sea. The glaciers that do not reach the coast often terminate in lobes within inland valleys, or sometimes coalesce and form piedmont glaciers on the adjoining lowlands (an excellent example of this is located south of Cape Bowen on northeastern Baffin Island). In the western part, glaciers form either very short projections on the periphery of the ice caps, or very long ones, such as those that terminate in Erik Fiord and Oliver Sound. The

terrain, where free of ice, is dissected by numerous glacial valleys floored with drift or alluvium. In the eastern part, these valleys are often completely or partly filled with lobes of ice fed by steep outlet glaciers that descend from the ice caps. The valley joining the upper reaches of Icy Arm and Coutts Inlet is a good example. These lobes of ice often dam lakes. Examples of this can be found at the head of Icy Arm, Coutts Inlet, North Arm, and Erik Harbour. Inland, the highland retains, despite its extensive dissection, a plateau character. Steep slopes occur mainly near the coast, although near sea level, steepness is often reduced by the presence of talus of colluvial debris and terraces cut into lateral moraines.

West of Oliver Sound, shorelines are mostly steep and rocky. They are often fringed by talus of colluvium and sometimes by narrow boulder beaches. The occurrence of gently sloping, unconsolidated backshores with continuous gravel beaches is restricted to the head of fiords and bays (White Bay, Tay Sound, Paquet Bay and Oliver Sound). East of Mittimatalik to Cape Bowen, steep and rocky backshores fringed by continuous boulder and gravel beaches alternate with moderately and gently sloping backshores with continuous gravel beaches. This last type of shoreline occurs most frequently in front of valleys that open to the sea. The shorelines of North Arm, Coutts Inlet, Icy Arm and Quernbiter Fiord consist of steep rocky backshores with or without colluvium, and narrow continuous boulder beaches. These alternate, along the fiords, with glacial ice fronts and with gently sloping, unconsolidated backshores with continuous gravel and sand beaches at the heads of fiords.

1.1.2.4.2 Buchan Highland

The area lying between Quernbiter and Dexterity fiords (Figure 5, Subdivision 5) is characterized by irregular but level-topped peninsulas and islands with steep coastal cliffs separated by closely-spaced, branching fiords (NTS Map 37H: Buchan Gulf). The area is also notable for the absence of lowlands, except for a very minor strip of coastal plain on the northeastern side of Dexterity Island. Several permanent ice fields are present in the central part of peninsulas, and they feed numerous narrow glaciers that either reach the sea or terminate in inland valleys. Several types of valleys are present: a few are broad, U-shaped, floored with drift, and lead far inland or cut across peninsulas; many are narrow, usually steep, and carry ice from ice fields to the sea; and many V-shaped ones breach the fiord walls. Summit elevations attain 1,200 m but most of the flat tops lie at about 1,000 m.

Shorelines along the fiords and around the islands are generally steep and rocky, often with talus of colluvium and narrow continuous beaches of boulders. Sections of shorelines that occur in front of valleys and at the head of fiords consist usually of moderately or gently sloping, unconsolidated backshores fringed by continuous gravel and sand beaches. Similar types of shorelines occur along the northeastern side of Dexterity Island where they fringe a two km-wide strip of coastal plain. Glacial ice also reaches the sea and, although relatively abundant, these ice fronts are invariably narrow and make up only a small proportion of the coastlines.

1.1.2.4.3 Clyde Highland

Between Dexterity and Tingin fiords (approximately 69° N), eastern Baffin Island is characterized by the presence of many steep-walled fiords that extend inland for long distances, and divide the area into roughly rectangular peninsulas (Figure 5, Subdivision 6, NTS Maps 37H: Buchan Gulf, 37E: Conn Lake, 27F: Clyde River, and 27C: McBeth Fiord). Near the coast, the terrain is deeply dissected and alpine in character, with valley glaciers descending from cirques or small ice fields, and mountains with sharp horns and arêtes. Further inland, larger ice caps are pierced by sharp nunataks that reach elevations of 1,500 to 1,800 m. Despite the deep dissection, there is a general accordance of peaks at an elevation of about 1,800 m. This suggests that there may once have been a well-developed upland surface at that altitude. The highest elevations occur in a belt parallel to the coast. Major fiords transect the highland area completely and a few extend into the lower plateau area to the west. These fiords are almost invariably steep-walled with unstable talus at sea level. As in the Buchan Gulf area, tributary valleys are either 1) occupied by valley glaciers that cut the fiord walls or that terminate a short distance inland, often against terminal moraines; 2) V-shaped stream valleys; or 3) broad and U-shaped, floored with alluvium, and leading far inland or cutting across peninsulas (as between Eglinton and Sam Ford fiords and the valleys leading east from Dexterity Fiord). Many rivers drain the plateau to the west of the highland and enter at the head of major fiords. There, they have cut gorges into the headwalls of the fiords and have deposited alluvium at times of higher sea level. These deposits are now being dissected into terraces. Small, discontinuous high-level benches occur occasionally along some of the fiords, and the limited areas of level land are covered by a bouldery drift.

As in the Buchan Gulf area, shorelines are mostly steep and rocky, often with talus of colluvial debris and narrow, boulder beaches. At the intersections of tributary valleys and fiords, and at the head of fiords, the gently and moderately sloping backshores are fringed by gravel and sand beaches.

1.1.2.4.4 Penny Highland

The Penny Highland is a deeply dissected area of mountains that cover over 3,000 km^2 on southeastern Baffin Island (Figure 5, Subdivision 7, NTS Maps 26N, 26O: Nedlukseak Fiord, 26P: Okoa Bay, 26J: Clearwater Fiord, 26I: Pangnirtung, and 16K & L: Cape Dyer). Most of the mountains are steep-sided, but their summits can include flat or gently sloping surfaces. Numerous long, narrow, and sometimes discontinuous valleys cross the area; the majority are filled with glaciers, but two major ones (Pangnirtung and Kingnait) are free of ice and divide the highland into three parts.

Pangnirtung Pass is a two km-wide valley that extends between Cumberland Sound and Davis Strait. To the northwest, a large part of the area is covered by the Penny Ice Cap whose summits reach 1,980 and 2,050 m

(Baird, et al., 1953). The ice cap is dome-shaped and rises from about 600 m in the northwest to over 1800 m in the southeast. Long outlet glaciers drain the eastern half of the ice cap through deeply incised valleys. Two of these reach the sea in the northeast, and some like the Coronation Glacier are over 30 km long and three km wide. Between Pangnirtung and Kingnait passes, and east of Kingnait Pass, smaller permanent ice fields, cirques, and valley glaciers are present, and the topography is one of complex, dissected mountains.

The Penny Highland appears to have been produced by the erosion of a rolling upland surface. Remnants of this surface are sometimes free of surficial deposits, but in general, they are covered with felsenmeer. In deep, ice-free valleys, a variety of terrains developed on glacial deposits. Sand and silt terraces, deltas, large areas of bouldery ground moraines, and lakes are frequent. Moraine ridges often contain ground ice, and extensive marshes are present where drainage has been blocked by these deposits.

From Padloping Island to Cape Dyer, steep, rocky coastlines predominate. These are sometimes fringed by talus of colluvial debris and continuous or pocket beaches of boulder and gravel. Where valleys reach the coast, backshores are often moderately or gently sloping and unconsolidated, except north of Cape Dyer where they are locally steep.

1.1.3 Baffin Upland

On central Baffin Island, a wide belt of plateaus and hills extends from Milne Inlet to Davis Strait and Cumberland Sound (Figure 3, Division 3). The area is characterized by a general uniformity in topography and by the presence of upland surfaces even where the terrain is dissected by deep valleys. East and north of Steensby Inlet, the upland is bounded by fault-line scarps. Southeast of the inlet to about 69°N, it rises directly from Foxe Basin. South of 69°N, the western limit of the upland swings a considerable distance inland.

A well-developed plain lying at an approximate elevation of 400 m extends along the western margin of the upland from the head of Steensby Inlet to about 68°N. This plain is bordered on the east by a belt of dissected plateaus partly covered by the Barnes Ice Cap. This belt forms a transition to the higher surfaces present in the eastern parts of the upland. North of Home Bay, the eastern boundary of the uplands is marked by a steep rise to the Davis Highlands.

The Barnes Ice Cap and smaller independent ice caps are remnants of ice which, during the maximum of the last glaciation, completely covered the upland of Baffin Island. Ice appears to have flowed outward from a divide lying along the centre of Baffin Island, towards Foxe Basin in the southwest and through valleys and fiords to Baffin Bay and Davis Strait in the east and southeast. Glacial deposits occur frequently on the upland. Terminal moraines, ground moraines, and a bouldery or sandy till occur respectively northeast, south and west, and northwest of the Barnes Ice Cap. Most

valleys include some glaciofluvial or glaciolacustrine deposits. Near the ice cap, several large marginal lakes, of which Conn and Bieler are the largest, occupy the head of valleys that formerly drained ice into Foxe Basin. Strandlines north of the ice cap indicate that it held back similar lakes before shrinking to its present size.

West and northwest of the Penny Highland, the upland consists of a relatively little-dissected, rolling terrain, lying at a mean elevation of about 600 m and rising gently from west to east. Immediately northeast and southeast of the highland, a zone of upland 20 to 30 km wide reaches elevations of 700 to 900 m. Elsewhere in the southern part of Baffin Island, the upland surface is rolling and hilly, with a local relief of 200 to 360 m. Slopes are usually smooth and rarely steep, except where the terrain is incised by fiords. Small areas of alpine mountains with cirque glaciers and small ice fields reach elevations of 1500 to 1800 m west of Cumberland Fiord. The subdivisions making up the Baffin Upland are illustrated in Figure 6.

1.1.3.1 Tay Area

A wedge-shaped area of Paleozoic sedimentary rocks extends southeasterly from Milne Inlet to a point beyond the head of Paquet Bay (Figure 6, Subdivision 1). The area is approximately 25 km wide by 120 km long and lies between the Pond Highland to the north and the Kragg and Rowley plateaus to the south (NTS Maps 48A: Milne Inlet, 38B: Pond Inlet, and 37G: Icebound Lakes). Near the southern boundary of the area, south and west of Tay Sound, a series of dikes associated with strike faulting stand up as low ridges, sometimes with vertical walls, in the otherwise gently undulating terrain. South of the head of Paquet Bay, Tay Sound and Eskimo Inlet, the sedimentary rocks form broad ridges. Hillsides are covered with a bouldery drift that is usually thicker at lower elevations. Till and alluvium occur in the broad valleys that are present. These valleys often include lakes and sometimes bogs. In general, drainage is shallowly incised into the surficial sediments.

The surface of the Tay area rises towards the north where it merges with the Pond Highland. At the junction, the erosional surface present on the highland transects the sedimentary rocks of the Tay area. Streams leading down to inlets are usually deeply incised and steep slopes occur at the shorelines.

The upper reaches of Eskimo Inlet, White Bay, Tay Sound and Paquet Bay have shorelines that consist of steep, rocky backshores with continuous gravel and boulder beaches, except at the heads of inlets and in front of valleys. There, backshores are often gently sloping, unconsolidated, and fringed by continuous beaches of gravel or sand. Major streams have formed substantial deltas where they enter the inlets. The lower part of the valley entering at the head of Paquet Bay has extensive deposits of surficial sediments, several terminal and ground moraines, a few bogs, and areas of dissected outwash fans and raised lacustrine or marine beaches.

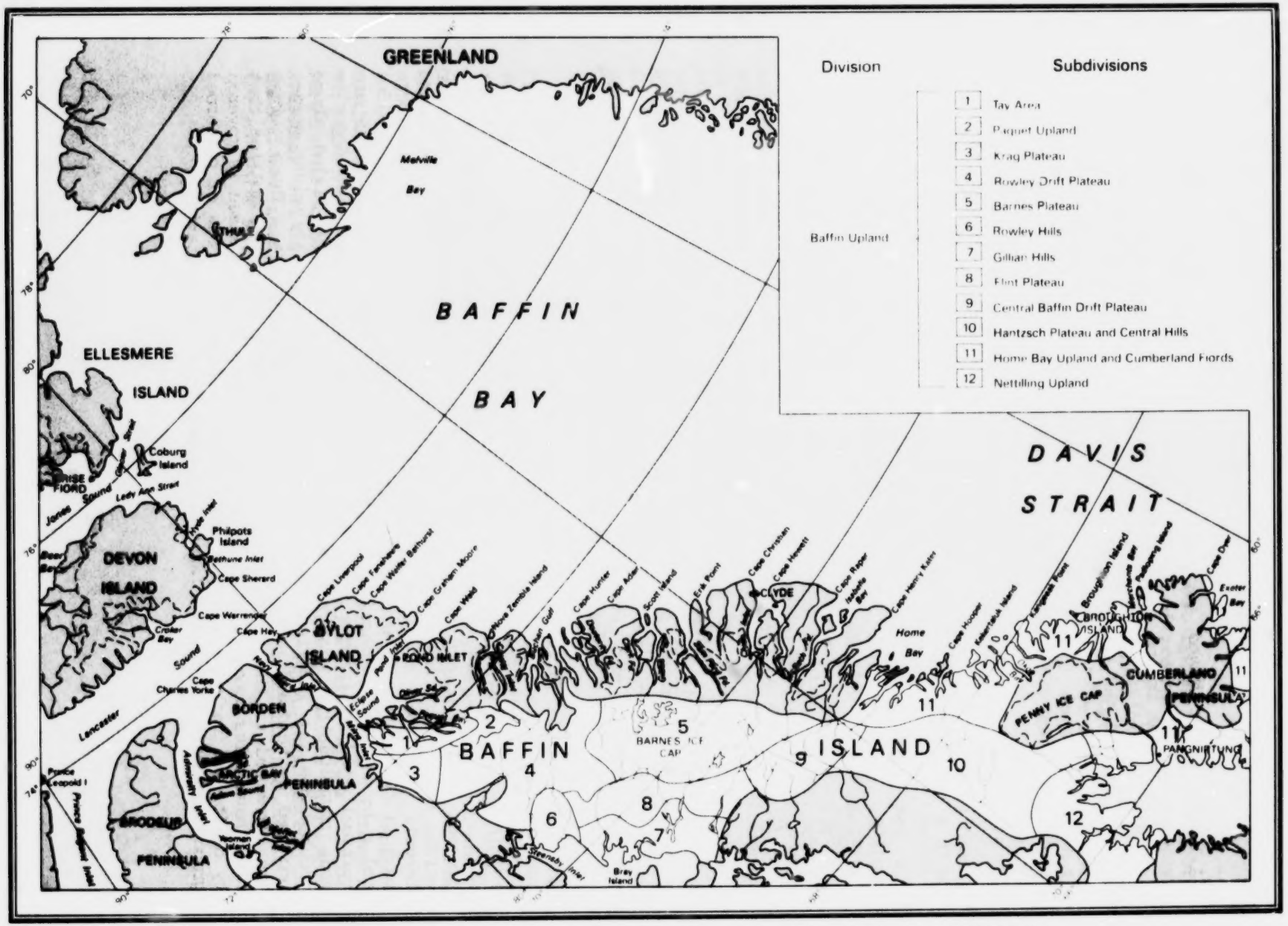


Figure 6 Physiographic Subdivisions of the Baffin Upland, after McGill University, 1963b and c.

1.1.3.2 Paquet Upland

The Paquet Upland forms a southeasterly continuation of the Tay area (Figure 6, Subdivision 2). It covers about 1,000 km^2 and extends between Paquet Bay and the head of North Arm (NTS Map 37G: Icebound Lakes). The upland area is underlain by Precambrian gneisses and metasediments that have a well developed west-northwest east-southeast structural trend. In the west, these Precambrian rocks are overlain by the sedimentary strata of the Tay area.

The upland is divided into two main parts by a structurally-controlled broad valley that joins the head of Paquet Bay and that of Coutts Inlet, with a branch leading to the head of North Arm. To the south of this main valley, the land rises steeply to the Rowley Plateau and is cut by a series of deep glacial valleys that intercept the structural trend at right angles. These valleys are invariably floored with alluvium, and a few contain lakes. North of the main valley, the upland has been dissected into roughly rectangular blocks of variable dimensions, and with summits lying at an altitude of about 600 to 760 m. Three types of valleys occur between these blocks. Some are cut 60 to 100 m below the surface of the upland and are floored with a coarse deposit of boulders, sand, and silt. They are generally conspicuous for their lack of vegetation in spite of their mature form. This lack of vegetation may indicate that ice caps have retreated in the recent past to their present position north of the area. Some of these valleys have upstream sections that have been rejuvenated and resemble gorges. In the eastern part of the upland, valleys are much narrower, and those draining into North Arm and Coutts Inlet have ice-dammed lakes at their mouths. The level of these lakes appears to fluctuate periodically as the ice dams sometimes give way during the summer.

1.1.3.3 Krag Plateau

South of the Tay Area lies a triangular region of 1,600 km^2 of rolling interfluves separated by deep valleys (Figure 6, Subdivision 3). This region, the Krag Plateau, is composed of Archean metamorphic and igneous rocks that are overlain in the northern part by the sedimentary strata of the Tay Area (NTS Maps 47H: Phillips Creek, and 37G: Icebound Lakes). Two types of valleys drain into Tay Sound and divide the northern part of the plateau into blocks. Some are mature with gently sloping sides and a broad, drift-covered floor, sometimes with small lakes. Others are vertically-walled troughs floored with alluvium or stagnant ice deposits and, occasionally, long narrow lakes. The ice that once filled these troughs has not modified their shape much; they are really river gorges produced by headward erosion of the northwestward drainage. These gorges are connected to the interfluve surfaces by short and steep gullies that often have talus cones at their base. The surface of the plateau is covered by drift, and includes numerous small lakes, substantial areas of patterned ground southeast of Icebound Lakes, and areas devoid of vegetation.

The Krag Plateau is bounded on the northwest by the upper reach of Milne Inlet. There, shorelines consist mainly of moderately and steeply sloping, rocky backshores, sometimes with colluvium, and often fringed by continuous gravel, sand or boulder beaches. Where valleys enter the inlet, shorelines usually consist of gently sloping, unconsolidated backshores fringed by continuous gravel or sand beaches. South of Tikerkdjuak Mountain and opposite Koluktoo Bay, the shoreline is relatively unusual: it includes long sections of moderately sloping, unconsolidated backshores, fringed by continuous beaches of silt and wide intertidal zones of silt and sand. These occur in front of extensive glaciolacustrine deposits of silt. A short section of shoreline opposite Poirier Island includes a boulder barricade; an unusual feature in the northern part of the study area.

1.1.3.4 Rowley Drift Plateau

The Rowley Drift Plateau covers more than 12,000 km^2 on central Baffin Island (Figure 6, Subdivision 4). It is bounded in the northwest by the Paquet Upland, and in the east, it interfingers with the eastern Baffin Highlands. Fault-line scarps delimit its southern edge, and in the south, it is in contact with the lower terrain of the Flint Plateau (NTS Maps 37G: Icebound Lakes, 37H: Buchan Gulf, 37F: Steensby Inlet, and 37E: Conn Lake). Summit elevations lie at about 600 m throughout the area, and the surface is either gently rolling with a relief of about 60 m, or maturely dissected with a relief of about 180 m. Occasionally, along very large rivers, local relief exceeds 300 m. Drainage appears to be structurally-controlled in spite of the drift that covers most of the plateau. In the northern part, the major trend in drainage is southeasterly to Steensby Inlet. In the southern part, the main streams flow southwestward, and many include rock basin lakes. An excellent example of this type of stream is Rowley River.

Several small, irregular ice caps, ranging in size from about one to 75 km^2, occur in the northernmost part of the plateau, near the junction with the Paquet Upland. They are probably remnants of more extensive ice caps, as they are presently surrounded by areas of light-coloured, vegetation-free drift. A few of these light-coloured, vegetation-free areas occur near ice-free summits and may indicate the presence of former ice fields.

Drift cover varies throughout the area. In the shallowly-incised northwestern part, a sandy drift obscures all bedrock. In the southwest, the cover is thin and infrequent, and numerous lakes appear to be structurally-controlled. Elsewhere, the surficial cover consists largely of boulders, and is generally thicker on summits than in valleys, except in their lower reaches where stagnant ice, lacustrine, and alluvial deposits are generally present.

1.1.3.5 Barnes Plateau

The central part of Baffin Island is occupied by a very large plateau that covers more than 28,000 km^2 (Figure 6, Subdivision 5). The centre of the plateau is occupied by the Barnes Ice Cap (Loken, 1966), itself covering more than 6,000 km^2 (NTS Maps 37E: Conn Lake, 37D: Lake Gillian, and 27C: McBeth Fiord). The boundaries of this plateau are somewhat arbitrary, as the transition to the Rowley Drift Plateau to the northeast and the Central Baffin Drift Plateau to the southwest is not marked by a topographic discontinuity. Differentiation between these areas is based on extent and freshness of glacial deposits.

Northeast of the Barnes Ice Cap, the Barnes Plateau reaches an altitude of about 600 m. Southwest of the ice cap, summits generally lie at an elevation of about 400 m. The ice cap itself reaches an altitude of 1,123 m, with a probable maximum ice thickness of about 450 m. Near its margins, the land elevations range from 400 to 580 m. The ice cap has a smooth dome shape with an almost circular outline in its southeastern part. Elsewhere, it has lobate margins that rest on land, sometimes against moraines and scree deposits, or into marginal lakes. Only one short and minor outlet glacier, the Lewis Glacier, is present at the northwestern edge of the ice cap. Along the margin, the ice is usually mantled with morainic material and slopes are as steep as 40°. Cliffs occur where the ice is undercut by marginal streams and where it calves into marginal lakes. On the ice cap, numerous stream channels develop into gorges during summer, otherwise there are no permanent crevasses transverse to the slope.

The regional slope in the vicinity of the Barnes Ice Cap is to the southwest. Consequently, the normal drainage of the valleys northeast of it is blocked and marginal lakes have developed. The largest ones, Conn and Bieler lakes, occupy a shallow, branching valley system and drain through the valleys of Tay and Erik rivers, into Gibbs Fiord. Southeast of Bieler Lake, local relief increases and isolated rounded hills are separated by structurally-controlled valleys. Southeast of the river draining to Walker Arm, and around the southeastern part of the ice cap to Flyway Lake, the terrain is rolling, with hills rising 90 to 120 m above the general level. The drift present on their summit is often of a light colour and free of vegetation. Marginal drainage channels are often present on sides of hills that face the ice cap. On the opposite sides, drainage channels radiate downslope. In some low areas, drainage is blocked by moraines, and bogs have developed. Near the ice cap, the unvegetated moraines are poorly drained and show evidence of slumping and heaving. Southeast of the ice cap, striaes and the slope and orientation of roches moutonnées, indicate that the last ice movement was to the northeast. Rivers draining marginal lakes are broad and shallow on the upland, but they narrow and become trough-like at the head of fiords in the highlands. At the southeastern extremity of the ice cap, Generator Lake is repeatedly crossed, in its northern part, by narrow morainic ridges. It drains through numerous rapids and smaller lakes into Clyde Inlet.

The area south of the ice cap, between Flyway Lake and Drewry River, is characterized by numerous glacial deposits and by a rolling surface lying at a mean elevation of 400 to 450 m. This surface is cut by broad, shallow, structurally-controlled valleys that drain meltwater southwestward from the ice cap. Immediately west of Flyway Lake, the ice has left a well defined terminal moraine about five km from its present position, probably during the very recent past 100 years. Soil polygons occur west and south of Flyway Lake. Substantial terminal moraines can be traced as far as 30 km from the southern margin of the ice cap. Blanchfield Lake, a marginal lake on the southern part of the ice cap, owes its origin to a lobe of ice that blocks the upper reach of MacDonald River. There, the cover of ground moraine is continuous, drainage is poor, and bogs are frequent. Between MacDonald and Drewry rivers, this ground moraine is locally very thin, and fails to completely cover the underlying scoured bedrock. Elsewhere along the upper part of MacDonald River, the drift cover is thick, drainage is poor, and bogs have developed.

Between Drewry River and the northwestern tip of the ice cap, the terrain consists of a dissected plateau with summits at an elevation of about 570 m and a relief of about 150 m. All major rivers drain to the west or southwest. The upper parts of valleys often include arcuate moraines that indicate former positions of ice lobes (Ives and Andrews, 1963). Elsewhere along their courses, valleys are floored with terraced outwash material that becomes progressively thicker near the ice cap, and rivers have numerous lake-like expansions.

The area adjacent to the northern tip of the ice cap is the most dissected part of the Barnes Plateau. There, summits lie at an elevation of about 600 m, and numerous southward draining valleys lie 120 to 150 m below the level of the interfluves. The interfluves themselves have been dissected into rounded hills by tributary valleys in which bedrock is often exposed. Strandlines occur frequently in major valleys, and mark the edges of former ice-dammed lakes. In the field, these strandlines are sites of beach accumulations, transitions in the degree of vegetation cover, and absence of fine particles in the drift. Many valleys also include transverse ridges that are probably remnants of terminal moraines. These ridges form obstacles to the drainage, and areas of bogs and lakes have developed between them.

1.1.3.6 Rowley Hills

An oval area of dissected terrain rises from the eastern shore of Steensby Inlet and extends approximately 70 km northeast (Figure 6, Subdivision 6). This area, the Rowley Hills, has a surface that reaches an elevation of about 360 m in its eastern part, and an average local relief of about 150 m (NTS Map 37F: Steensby Inlet). The area is most dissected near the inlet. There, numerous narrow lakes and river valleys appear to be structurally-controlled. Further inland, dissection decreases and the interfluves become larger. The interfluves are usually covered with a thin

layer of bouldery drift, and deeper valleys often include deposits of sand and gravel, especially near their mouths. The mouth of Rowley River has an extensive area of silt and alluvium deposits on which large marshes have developed. The northeastern shore of Steensby Inlet is very indented with numerous valleys leading far inland.

1.1.3.7 Gillian Hills

South of Rowley Hills, a belt of very dissected terrain 30 to 50 km wide rises from the eastern coast of Foxe Basin and extends to the Flint Plateau (Figure 6, Subdivision 7). The area is characterized by a deep dissection: elevations seldom exceed 270 m but the local relief is sometimes greater than 200 m (NTS Maps 37F: Steensby Inlet, 37C: Koch Island, and 37D: Lake Gillian). Glacial deposits have formed numerous interconnected lakes. Many of these lakes are eight to ten km long and most are joined by short, rapidly flowing rivers. In the main valleys, rugged hills, sometimes with steep sides and a thin, discontinuous cover of bouldery drift, rise 100 to 150 m above the floor. Many valleys, especially deeper ones, are comprised of eskers, terraces and stagnant-ice deposits of sand and gravel. Rivers draining the Barnes Ice Cap appear to carry much silt.

Along the coast north of Baird Peninsula, outwash deposits have been reworked into raised beaches within a three km-wide strip. Raised beaches occur to a maximum elevation of 90 m. Opposite Bray Island, this coastal strip widens and includes very wide (one to two km) intertidal zones. From there to a point beyond Grant-Suttie Bay, shorelines are indented and steep bluffs alternate with narrow pockets of alluvium. In Grant-Suttie Bay, raised beaches occur to an elevation of 96 m. The coast of Steensby Inlet is very straight along its southeastern part, and consists entirely of steep bluffs that rise 75 m or more directly from the sea.

1.1.3.8 Flint Plateau

The Gillian Hills and the Barnes Plateau are separated by an elongated belt of variable width, the Flint Plateau (Figure 6, Subdivision 8). Its longest dimension, about 200 km, is oriented northwest to southeast and its width varies between 20 and 50 km (NTS Maps 37F: Steensby Inlet, 37E: Conn Lake, and 37D: Lake Gillian). In the northern part, the surface of the plateau reaches an elevation of about 390 m, and it eventually merges with the upland surface present south of the Barnes Ice Cap. South and west, the general elevation of the terrain declines to about 255 m, and its surface continues into the area of Gillian Hills. The general drainage is to the southwest along a relatively well-defined structural trend. Considerable relief, 150 to 350 m, is present along some of the major river valleys, such as those that include Flint Lake and the one north of Isortoq Fiord. Other river valleys, like MacDonald and Drewry, have a mature, open form but have tributary valleys that are sometimes deeply incised. Interfluves are characterized by a very small, local relief. Numerous shallow lakes appear to occupy structurally-controlled, preglacial valleys that have been blocked by deposits of drift.

Except for an area between Grant-Suttie Bay and the Barnes Ice Cap where the drift is thick, the surficial cover on the plateau is generally thin and discontinuous. In the area of thick cover, the surface is crossed by end moraines oriented parallel to the edge of the Barnes Ice Cap. The broad valleys that are present are usually floored with sand and gravel deposits and often include numerous lakes.

1.1.3.9 Central Baffin Drift Plateau

South of the Barnes Ice Cap at the latitude of Home Bay, there is a roughly rectangular area that covers about 6,300 km^2 and is characterized by flat or gently undulating interfluves (Figure 6, Subdivision 9). Elevations are greatest in the northeast next to the Baffin Highlands, and the general level declines southward to about 480 m, and westward to about 390 m (NTS Maps 37D: Lake Gillian, 27C: McBeth Fiord, 37A: Foley Island, and 27B: Ekalugad Fiord). Drainage, except in the easternmost part of the area, is controlled by three structural elements of the underlying bedrock: the east-west strike of the highly folded metamorphic rocks, and systems of northwest and northeast trending joints. In the eastern part of the plateau, major rivers flow south through a series of lake-like expansions. In general, rivers are shallowly incised, and the resulting relief is very small, about 30 m. Drift cover is generally thick, except in the southwestern part of the plateau. Large areas of soil polygons and ground moraines are present, and a series of terminal moraines lie in the south-central part.

1.1.3.10 Hantzsch Plateau and Central Hills

A rectangular area joins the Central Baffin Drift Plateau in the north to the Nettilling Upland in the south (Figure 6, Subdivision 10). The area is approximately 200 km wide and 650 km long, and has a mean elevation of about 420 m (NTS Maps 27B: Ekalugad Fiord, 26M: Hantzsch River, 26N, 26O: Nedlukseak Fiord, 26K: Amittok Lake, and 26J: Clearwater Fiord). In the northeastern part, the surface has a lower elevation and is less dissected than in the adjacent Home Bay Upland. On the western side, elevations decrease to about 360 m. In the southern part, elevations increase in the area of the Central Hills. Locally, broad valleys 100 to 180 m deep divide the upland into blocks of rolling hills. Hill tops have a discontinuous bouldery cover, and valleys often include ground moraines, stagnant-ice deposits, and numerous lakes. To the north of Hantzsch River, valleys are cut 60 to 120 m below the elevation of the interfluves. Drainage is largely controlled by bedrock structures and is composed of numerous lakes interconnected by short sections of streams. South and west of Hantzsch River, the plateau is lightly dissected, relief is shallow, about 60 to 90 m, and interfluves consist of rounded hills. Numerous lakes of various sizes and shapes are present, especially west of Hantzsch River. Many of these lakes occupy rock basins, and those that occur in areas of relatively deep surficial deposits are generally shallow and bordered by swamps. Elsewhere, the interfluves are free of vegetation.

The Central Hills occupy about the southern half of this subdivision. In this part, elevations rise gently from west to east and have a mean value of about 645 m. As in the north, drainage is controlled by bedrock structures, but to a lesser degree than in adjoining areas.

West of Nettilling River, the terrain is transitional to the lowland on the west, and drainage is to the west. Valleys are shallow but many have steep sides. Small lakes are abundant.

To the east of middle Nettilling River and its broad, sand-filled and terraced valley, the land rises gently towards the Penny Highland. The area is crossed by the upper Nettilling River and its tributary streams. The valleys in which these flow are usually broad, shallow, and floored with sand and silt.

South of the upper Nettilling River, the rolling hilly interfluves merge into a plateau about 20 km west of the Penny Ice Cap. At that point, surficial deposits change from a bouldery drift to a series of morainic ridges made up of angular boulders and sand. Large sandy areas and shallow lakes are abundant, and it is clear that the terrain has only recently been free of ice.

1.1.3.11 Home Bay Upland and Cumberland Fiords

Beginning south of Pitchforth Fiord (at approximately 68°N of the eastern coast of Baffin Island), there is an elongated, blocky plateau dissected by deep valleys and fiords (Figure 6, Subdivision 11). Near the coast, dissection is more pronounced and the remnants of the plateau surface are smaller and more frequent. Further inland, the rolling plateau surface has a relief of about 300 m, and it reaches elevations of about 600 m (NTS Maps 27B: Ekalugad Fiord, 27A: Home Bay, 26O: Nedlukseak Fiord, 26P: Okoa Bay, 16M & N: Padloping Island, 26J: Clearwater Fiord, 26I: Pangnirtung Fiord, 16K & L: Cape Dyer, 26H: Abraham Bay, and 16E & D: Hoare Bay). Higher parts of the plateau include small highland glaciers; elsewhere, the surface is generally covered with a bouldery drift. Most valleys appear to be structurally-controlled, and all larger ones have been glaciated and overdeepened in their lower sections to produce fiords. Further inland, larger valleys are floored with sand and gravel, and have streams that are often braided and have deltas at their mouths. Valleys at higher levels frequently include lakes that drain to the fiords through steep and narrow streams.

Northeast, southeast and southwest of the Penny Highland, the upland forms elongated zones, approximately 30 km wide, with elevations that reach 800 m. The terrain is rolling and hilly and has a local relief of up to 360 m. Slopes are usually smooth, and except for fiords, they are seldom steep. Small areas of the Penny Highland have been preserved on the upland, and form small alpine mountains with cirque glaciers, local ice fields, and elevations that reach 1,800 m. Smaller valleys are often V-shaped with

steep sides, but most are shallowly incised. Many enter main valleys by gullies and falls. Larger valleys are U-shaped, several kilometres wide and up to 30 km long. They often include morainic ridges and outwash deposits. Other valleys have been incised below present sea level, and their submergence has produced fiords.

Northwest of the Penny Highland, the area is deeply incised, little of the upland remains, and the whole region has been cut into blocks by long fiords. The coast includes little flat ground, except on islands and at the headland of longer peninsulas.

South of Cape Hooper to Kangeeak Point, the mainland coastlines consist of steep, rocky backshores with continuous boulder beaches and colluvial talus. At the head of fiords and at the junctions of fiords and tributary valleys, shorelines consist of gently or moderately sloping, unconsolidated backshores fringed by continuous gravel and boulder beaches. Small islands oten have a steep, rocky shoreline either devoid of sediments or with some pocket beaches of gravel and boulders. Larger islands, like Kekertaluk, also comprise small areas of lowland that terminate into gently sloping, unconsolidated backshores and continuous boulder and gravel beaches.

Similar shoreline types are present south of Kangeeak Point with two exceptions. The northern part of the peninsula that includes Kangeeak Point and headlands located between that same peninsula and Maktak Fiord, include substantial sections of shorelines made up of steep, unconsolidated, eroding backshores fringed by continuous boulder or gravel beaches. Towards Cape Dyer, backshores become increasingly steeper and rocky. Valleys leading inland from the coast become progressively more narrow, but shorelines at their mouths nevertheless consist of unconsolidated backshores and beaches of coarse sediments.

Southeast of the Penny Highland, between Cape Mercy and Cape Dyer, the upland is penetrated by many long fiords that invariably have steep sides, especially in the vicinity of Cape Dyer. Towards the south, fiords become shorter and give way to wide-mouthed short inlets. Inland from the coast, the topography is hilly.

South of the Penny Highland and northwest of Pangnirtung Fiord, many overdeepened valleys have been filled with silt and sand from the Penny Ice Cap. Further south, Pangnirtung and Kingnait fiords penetrate far inland and into the highland. The northeastern coast of Cumberland Fiord consists of steep cliffs with few sediments.

1.1.3.12 Nettilling Upland

East of Nettilling Lake, the elevation of the lowland increases and valleys become more dissected towards Cumberland Sound. The resulting dissected plateau area, the Nettilling Upland, covers about 4,000 km^2 of Baffin Island northwest of the head of Cumberland Fiord (Figure 6, Subdivision 12). The upland is dominated by linear scarps that form the

sides of numerous islands, inlets, lakes, and control drainage (NTS Maps 26K: Amittok Lake, 26F: McKean River, and 26G: Irvine Inlet). Inland, the plateau has an undulating surface at a mean elevation of about 250 m. Strong structural lineaments enhanced by erosion have produced straight, often steep-sided valleys of variable width. Drowning of the plateau at the head of Cumberland Sound has formed deep inlets and many isolated islands. Near the northern limit of the upland, the transition to the Central Hills is gradational, and in this area, dissection is less severe.

The surface of the plateau includes rock outcrops, rock basin lakes and some areas of boulder deposits. Surficial deposits, mostly boulders in a sandy matrix, are concentrated along valley floors, and often form relatively level areas partially covered with tundra vegetation. Some valleys include deposits of sand and silt at their mouths. These types of deposits occur more frequently in valleys that have streams that originate at the Penny Ice Cap.

1.1.4 Hall Upland

Three peninsulas extend southeasterly from Baffin Island (Figure 1). The central one, Hall Peninsula, lies between Cumberland Sound and Frobisher Bay, and forms part of the Hall Upland (NTS Maps 26F: McKeand River, 26G: Irvine Inlet, 26B: Chidliak Bay, and 26A: Leybourne Islands). Physiographic conditions vary widely within this upland, but transitions between units are generally gradational.

The core of the upland is characterized by high-level erosional surfaces preserved as a lightly-dissected plateau. Elevations increase towards the southeast where they reach 900 m. To the northwest, they decrease and eventually merge with the Nettilling Upland at the head of Cumberland Sound. The core area has a monotonous, gently rolling, lake-free, unvegetated surface with few deep valleys. In the southeast, low domes rise above the general surface and a few are covered with permanent ice fields. The northwestward decline in elevation takes place through a series of steps rather than a continuous slope. In the highest sections to the south, a few broad valleys lie about 45 m below the surface of the plateau. They are generally filled with sinuous, branched, shallow lakes. The eastern and southern sides of this part of the plateau are crossed by vertically-sided trenches occupied by rivers.

The Hall Plateau, the core of the Hall Upland, is almost completely surrounded by a rim of deeply dissected terrain. Part of this rim is present in the study area (Figure 3, Division 4). North of Chidliak Bay, the upland surface gives way to a lower erosional level at an average altitude of 450 m. Isolated hills and broad open valleys are present near Cumberland Sound. There, coastlines are characterized by numerous bays and low, rocky islands. Further inland, rivers are deeply entrenched and the interfluves are themselves dissected by shallow but steep-sided valleys. South of Chidliak Bay, the average level of the terrain rises to the edge of Cumberland Sound. In this area, cliffs over 600 m high, capped with permanent icefields, form the coastline.

1.2 The Borderlands

Like the Shield, the Borderlands can be divided physiographically into regions, divisions, and smaller units. Three major divisions of the Borderlands are present in the Arctic (Figure 1): the Arctic Coastal Plains, the Innuitian Region, and the Arctic Lowlands. This last region, the Arctic Lowlands, covers most of Banks, Victoria, King William, Prince of Wales, and Somerset islands; and within the study area, it extends onto southern Ellesmere Island, western Devon Island, Borden Peninsula, Brodeur Peninsula, and over a large part of Foxe Basin (Bostock, 1976). Two major divisions, the Lancaster Plateau and Foxe Plain, make up the part of the Arctic Lowlands that is present within the study area (Figure 7).

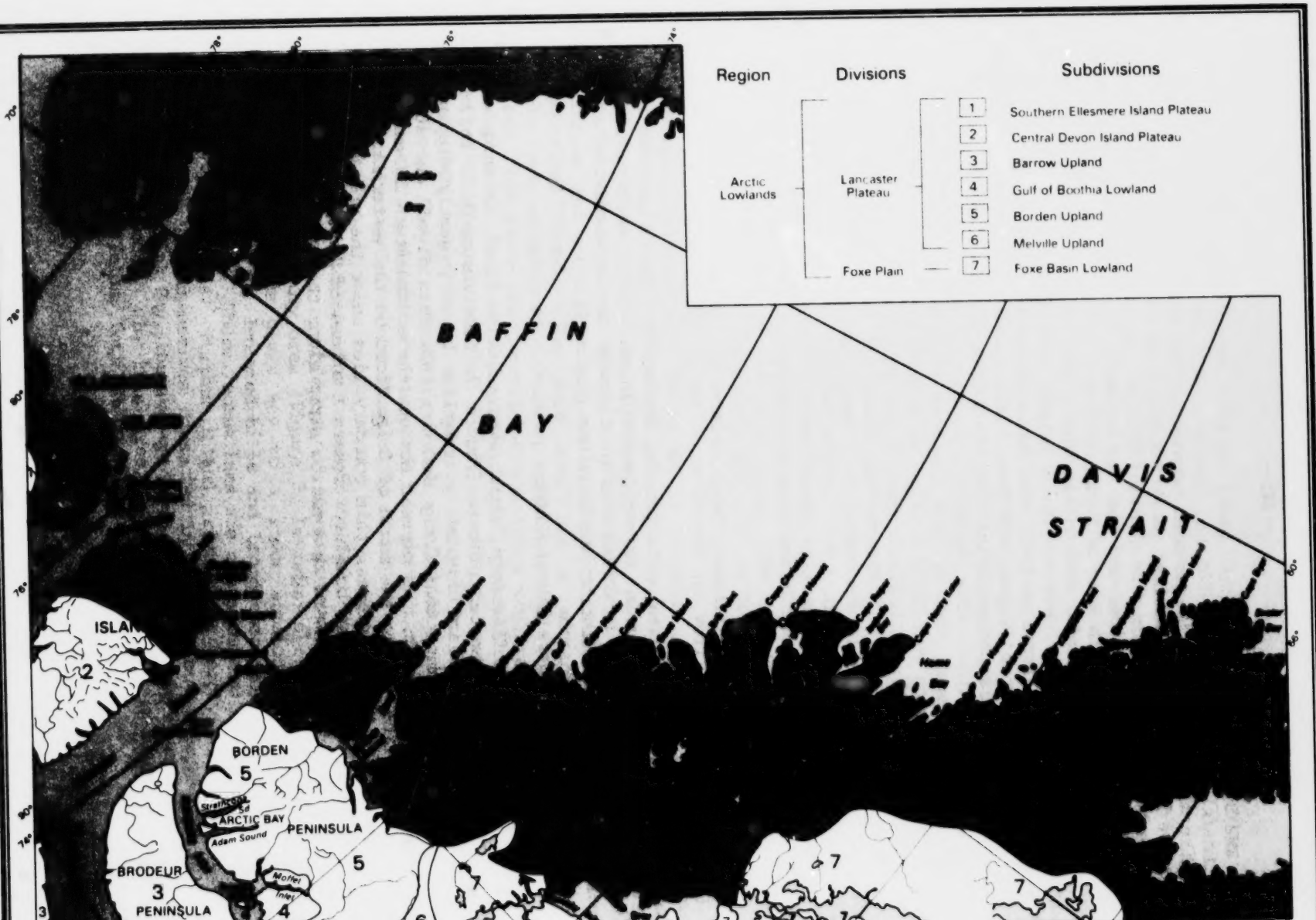
1.2.1 Lancaster Plateau

The part of the Arctic Lowlands east of the Boothia Arch (an elongated, north-south trending area of Precambrian rocks on southwestern Somerset Island and central Boothia Peninsula) has been named the Jones-Lancaster Plateau by Fortier et al., in 1954. Strata between the Boothia Arch and Admiralty Inlet have a basin-like distribution. Those on western Devon Island and southern Ellesmere Island form a broad syncline, and are probably part of this same basin.

On southern Ellesmere Island, the surface of the Lancaster Plateau slopes gently southward from an elevation of about 760 m, across central Devon Island, to an average elevation of 300 to 600 m on Somerset Island and Brodeur and Borden peninsulas. Its general uniformity appears undisturbed by Jones and Lancaster sounds, Prince Regent Inlet, and Admiralty Inlet. Within the study area, the Lancaster Plateau division can be subdivided and described in terms of six distinct subdivisions (Figure 7).

1.2.1.1 Southern Ellesmere Island Plateau

The southern part of Ellesmere Island, west of Grise Fiord, is occupied by a gently sloping, dissected plateau (Figure 7, Subdivision 1). It is approximately 50 km wide and completely underlain by very gently dipping sedimentary rocks (NTS Maps 49A: Craig Harbour, 49B: Baad Fiord, and 40D: Vendom Fiord). To the east, the plateau borders the southeastern Ellesmere Highland along a line joining the mouth of Grise Fiord to the western branch of Starnes Fiord. In many places, this boundary lies under the present ice cover. To the northeast, the boundary crosses the head of Makinson Inlet, and probably continues northward as a narrow strip along the western edge of the highland.


Region
Divisions
Subdivisions
Arctic Lowlands
Lancaster Plateau
Foxe Plain
1 Southern Ellesmere Island Plateau
2 Central Devon Island Plateau
3 Barrow Upland
4 Gulf of Boothia Lowland
5 Borden Upland
6 Melville Upland
7 Foxe Basin Lowland
BAFFIN BAY
DAVIS STRAIT
BORDEN PENINSULA
ARCTIC BAY
Adam Sound
BRODEUR PENINSULA
Moffet Inlet

North of Starnes Fiord, the surface of the plateau, where free of ice, reaches an altitude of more than 900 m. In its central part, an ice cap covers about 2,660 km^2 and reaches an altitude of more than 1,500 m (Sharp, 1956). The distribution and altitude of nunataks suggest that the ice cap rests on a relatively elevated and rounded part of the plateau. From the position of the ice cap, the general elevation decreases towards the north, south, and west. Along the southern coast of the island, the plateau lies mostly at an altitude of less than 750 m.

Compared to its counterpart on Devon Island, the Ellesmere Plateau is less uniform, more dissected, and its landscape shows more evidence of structural control. Erosion has emphasized features such as fault blocks and has produced discontinuous cuestas with gentle north slopes and short, steep, south-facing scarps; sub-rounded spurs and many shallow, V-shaped, high-level valleys. The original plateau surface is best preserved next to the highland between Starnes Fiord and Makinson Inlet, where interfluves have relatively smooth surfaces underlain by gently dipping strata.

Fiords dominate the topography of the plateau along the southern coast of Ellesmere Island. They are all several kilometres wide and from 28 to 50 km long. They are also invariably steep-walled and spaced with noticeable uniformity. Apart from fiords, other valleys are usually short, arranged in a dendritic pattern, and most have braided flood plains that are presently extending laterally, often by undercutting walls. A few major ones appear to be running parallel, and are possibly structurally-controlled. In most cases, the major valleys originate at the head of fiords, and cross the entire plateau area. Others, possibly eroded along faults, may predate the cutting of the fiords they cross.

The surface of the plateau is covered with a gravelly rubble. Locally, especially in the western part, the slope of the land has caused solifluction, mining of debris, and created soil stripes and rock streams. Elsewhere, mass movement has been very limited, and lithological differences in the underlying bedrock can be traced in the overlying rubble. There are very few lakes on the plateau, and those present appear to owe their origin to the blocking of valley streams by glacier tongues.

Ice fields and glaciers cover about 40 per cent of the surface of the plateau. It seems that at one time, the complete surface was under ice, but the effects of glaciation have been mild in the dominantly stream-cut terrain. Fiords along the southern coast must have been excavated by valley glaciers. The contrast between high-level, shallow, stream-cut valleys and fiords is due to the protective effect of the ice cover. The southern edge of the ice cap, with an outlet through Jones Sound, must have been the most active part, hence the development of fiords.

The southern coast of Ellesmere Island is bold, and consists to a large extent of cliffs with colluvial talus. There are no areas of extensive lowland, but short stretches of gently sloping backshores are present where

Figure 7 Physiographic Subdivisions of the Arctic Lowlands, after McGill University, 1963a, b, and c.

relatively deep valleys cross fiords. Deltaic flood plains at the heads of fiords are not as abundant as elsewhere in the eastern Arctic, but they do occur in front of a few major rivers. Areas of gently sloping backshores and of accumulations of colluvium often include well-developed raised beaches.

1.2.1.2 Central Devon Island Plateau

An extensive plateau lies west of the highland on central and western Devon Island (Figure 7, Subdivision 2). The area is part of the Jones-Lancaster Plateau (Fortier, 1957) and is characterized by a relatively featureless, well-preserved inland surface bordered by continuous coastal cliffs, fiords, and steep-walled bays (NTS Maps 58G: Baillie-Hamilton Island, 58H: Bear Bay West, 48G: Bear Bay East, 48H & 38G: Lady Ann Strait, 58E: Maxwell Bay, 48F: Powell Inlet, and 48E & 38F: Dundas Harbour). Where bedrock outcrops, as in the sides of fiords and bays, it consists of undeformed sedimentary strata. It appears that the plateau and perhaps most of the adjacent highland, developed as a single mature or old age erosional surface whose limits lay beyond the ones present on the island.

Elevations are highest in the eastern part, where the edge of the plateau is largely ice-covered and transitional to the highland. At the edge of the ice cap that covers most of eastern Devon Island, the plateau surface lies at an elevation of about 600 m. From there, it slopes gently westward and reaches an altitude of about 195 m on the west coast of the island. The paleozoic strata underlying the area have been well-preserved and dip west a bit more steeply than the plateau surface. The resulting effect is a series of north trending units intercepting the surface as bands that are progressively younger towards the west. Lithologies and structures have very little effect on topography, other than subtle differences in the texture of the drainage of adjacent bands.

Drainage and marine processes are the two main agents of erosion, and frost and glacial action have modified the terrain only in details. Streams are usually short, and flow independently to the coast from a divide located roughly along the middle of the island. Three main patterns of drainage are present: 1) subparallel, down along the slope of the surface; 2) structurally-controlled, mostly by joints in the underlying rocks; and 3) dendritic, as a result of headward erosion into uniform rocks. Most streams appear to have been cut since the surface was uplifted, and a few appear to coincide with tributaries to former main streams that may have flowed in the present position of Lancaster and Jones sounds (Fortier and Morley, 1956). Most streams are in the late youth stage, and have developed cliff-walled valleys tapering to ravines at their head. This is typical of stream erosion in a dry climate in mostly flat sedimentary strata. Many valleys have floors of fluviatile debris in the lower part of their courses. Some have built deltas as the sea level dropped; these deltas are presently being cut by their own streams. Most valleys have drowned mouths and must have been cut when the sea level was lower. On the north side of the island,

some streams have cut through all Paleozoic strata and are now widening their valleys and exposing a jointed, sediment-free Precambrian floor. Upstream, the gorge- like valleys narrow to V-shaped ravines and either end abruptly or meander without well-defined courses on the surface of the plateau.

Lakes are normally absent from the surface of the plateau, except in an area north of Maxwell Bay where numerous small ponds and depressions are present. In this area, the substrate comprises gypsiferous shales, gypsum and anhydride, which are mechanically flexible and are unlikely to control drainage structurally. As a result, progress of stream erosion has been irregular and few master valleys were able to develop. The ice that covered most of the island has generally protected the bedrock surface, except near outlets where its movements must have eroded its bed. Maxwell Bay and Bear Bay may have been such outlets, and erosion of a bedrock characterized by units, sometimes lenticular, of contrasting resistance, may have produced the topography present north of Maxwell Bay.

Southwest of the head of Bear Bay, the Haughton Dome consists of a core of gypsiferous rocks surrounded by steeply dipping, progressively younger beds. Recent studies (Frisch and Thorsteinsson, 1978; Robertson and Plant, 1981) have shown that the structure may be an astrobleme; that is, an ancient erosional structure produced by the impact of a cosmic body. In this area, stream erosion has produced a series of concentric valleys separated by curving ridges that are broken by narrow gaps.

The continuity of the southern coast of Devon Island is interrupted by several major fiords and bays, but has no promontories. Coastal cliffs are almost continuous and 150 to 300 m high. Fiords and valleys are spaced at roughly equal intervals and may have developed from short, consequent streams on the south-facing scarp that resulted from the uplift of the island. Except for Maxwell and Bear bays, there is little evidence of extensive erosion by ice, even in the U-shaped fiords and valleys along the southern coast. These could have been produced by stream erosion and frost action in jointed, flat-lying strata. The coastal cliffs are often protected by colluvium, but there are places where they reach the sea. Where this is the case, sea ice rather than wave action is the dominant erosion agent. On the cliffs that terminate directly into the sea, debris dislodged by frost action collects on an ice foot, and is often rafted away during spring. Erosion by sea ice may have formed the rock terraces that are locally present at sea level.

Along the western and northern coasts of the plateau on Devon Island, cliffs are intermittent. Where present, they are usually located at headlands and are separated by valleys. These valleys appear to be more structurally-controlled along the north coast than anywhere else. They usually terminate on a coastal plain made up of either unconsolidated sediments or very indented, low-lying bedrock. The surface of this bedrock is an exhumed unconformity between Precambrian and Paleozoic rocks.

1.2.1.3 Barrow Upland

Within the study area, the Barrow Upland covers a large part of Somerset Island and all of Brodeur Peninsula (Figure 7, Subdivision 3; NTS Maps 58D: Cape Clarence, 48C: Arctic Bay, 58A: Fitzgerald Bay, and 48B: Moffet Inlet).

On Somerset Island, the plateau has a gently undulating surface with an elevation of 345 to 390 m and local relief of less than 30 m. The area is underlain by sedimentary rocks, mostly limestones, conglomerates, sandstones, and shales. Streams flow across the surface in wide, shallow valleys. Rock outcrops on the plateau are rare. The surface is covered with a variable thickness of rubble composed of resistant weathered rock fragments, often cherty limestone, and larger boulders in a matrix of sand and clay. This surface deposit shows no sign of having been moved by ice. River gorges are deeply entrenched into the plateau surface and often have vertical sides. In their lower sections, major valleys are often open and have small, braided flood plains. In many locations, the plateau ends abruptly in coastal cliffs. The north coast of Somerset Island is broadly curved, convex to the north (McMillan, 1963). Garnier Bay and Cunningham Inlet are the only two major indentations. Four main deltas covered by raised beaches extend into Barrow Strait. Spits are present on the south ends of the flat-topped and cliffed Limestone Island and Prince Leopold Islands and indicate that there is a westward prevailing current in Barrow Strait adjacent to Somerset Island. High cliffs, mesa-like remnants of the plateau, rise from the shore around Port Leopold, Cape Admiral McClintock, Cape Rennell, near Cunningham Inlet and at Pressure Point. Elsewhere, the land rises gently from the shoreline and forms well-rounded, gentle hills several miles inland.

The shoreline is characterized by narrow but frequent bars and lagoons. Raised beaches are abundant up to 60 m above present sea level. The western shore of Prince Regent Inlet is relatively straight, and consists of almost continuous coastal cliffs. In the northern part, these cliffs are about 300 m high and their elevation declines to the south. Along the coast, the cliffs truncate the plateau surface and its incised drainage system. On this eastern part of the island, many streams appear to be headward portions of a drainage system whose main part must have been flowing in the area now occupied by Prince Regent Inlet. To the south, the plateau surface slopes down to the low, mud flat coastal strip that borders Creswell Bay.

Brodeur Peninsula is characterized by a plateau with a smooth surface bordered by continuous coastal cliffs and a dendritic-radial drainage system. The surface of the plateau lies at an altitude of about 540 m near Cape Crauford in the northeast and slopes gently and uniformly to an altitude of about 300 m in the southwest. Interfluves are smooth with a few low hills and from the ground, the peninsula appears as a flat and mostly featureless plain. Most of the surface consists of a gravelly rubble with little vegetation. Areas of felsenmeer are very limited, patterned ground is essentially absent, and solifluction lobes and soil stripes occur only on a few gentle slopes in the southern part of the peninsula.

Drainage on Brodeur Peninsula consists of a large number of small streams running east, northwest and west from a sinuous divide that parallels the long axis of the land. Stream pattern is dendritic, and it dissects the plateau into roughly equidimensional units. Major streams are sharply incised, and almost all streams are presently downcutting, and have smooth, longitudinal profiles. Most of the smaller tributaries are intermittent, and have cut ravines and gullies. Most streams have a V-shaped profile, and except in their lower courses, have no valley floor deposits or flood plain. Near the western coast of the peninsula, major streams have cut their walls to form a flat-floored valley, up to a few kilometres wide and seven km long. Occasionally, where more than one stream approaches or joins each other near the coast, their valleys merge and form sections of lowlands on the peninsula. These lowlands may be made up of distinct erosional surfaces lying at different elevations with respect to the present sea level.

Five small ice caps lie on the northeastern part of the plateau on Brodeur Peninsula. They are small, the largest being approximately 10 km wide, and all are thin. They all lie on interfluves, and a few cover the heads of stream gorges. A few have short tongues that extend along these gorges, but they are too short to be called outlet glaciers. None of these ice caps appear to have had an effect on the stream-cut landscape. The only evidence of glacial erosion is found between the present ice caps and the west coast of the peninsula in the vicinity of Jackson Inlet and Port Bowen. It consists of a different drainage texture, a gentler slope in the walls of gorges, and rounded irregularities on the surface of the plateau. It appears that all of Brodeur Peninsula was at one time ice-covered, but this ice cover did not significantly modify the underlying terrain.

Brodeur Peninsula is bounded by steep, continuous coastal cliffs often protected by colluvial talus and a narrow but continuous beach. On the western side of the peninsula, the coastline forms a smooth curve from Stanley Point to south of Port Neill. The cliffs are occasionally breached by deeply incised valleys and, locally, small points and capes consist of low spits or small deltas. There are no outstanding headlands. The coastlines appear to abruptly truncate the drainage system present on the plateau. Evidence indicates that the present drainage system must have developed on a land surface that extended in the area presently occupied by Prince Regent Inlet. Many valleys extending obliquely to the shore are headward branches of a dendritic system of which the trunk must have been present west of the actual coastline. Drainage patterns and stream incisions must have been established before the peninsula was reduced to its present shape.

Modern shoreline processes appear to be mostly depositional, as demonstrated by the building of spits such as Sargent Point, and deltas such as Cape York. Features along the coastline indicate that: 1) the valley gorges were cut before the coastal cliffs; 2) valleys were cut to about their present width when their floors were at or near present sea level; 3) coastal cliffs were cut back to their present position at a time when sea

level was similar to the present one; 4) submergence of as much as 60 m took place after the coastline reached its present position; and 5) present shorelines are relatively inactive, except where some rivers are building small deltas. Coastal cliffs are eroded mostly by frost action and beaches are reworked mostly by sea ice. Spits may build up on parts of the coast where major and presumably recurring leads develop in the pack ice.

1.2.1.4 Gulf of Boothia Lowland

Two areas of lowlands occur at the southeastern corner of Admiralty Inlet (Figure 7, Subdivision 4; NTS Maps 48B: Moffet Inlet and 47G: Berlinguet Inlet).

At the northern one, the hills and uplands are lower than further north, and on both sides of Moffet Inlet, the topography is more uniform than anywhere else on Borden Peninsula. The area is underlain by gneisses that were at one time overlain by sedimentary strata, except at two small locations in central Steensby Peninsula. Northeast of Moffet Inlet, hills reach an altitude of 150 m, and their elevation decreases to about 90 m southwest of the same inlet. In the southwestern part, the hills form an uneven terrain with small and steep ridges that alternate with depressions occupied by marine terraces and raised beaches. Bedrock is overlain by a discontinuous drift cover that becomes thicker to the southwest of Moffet Inlet.

On both sides of the eastern part of Berlinguet Inlet, the sedimentary rock cover has been eroded and the underlying granitic rocks form a flat and monotonous terrain. Some small areas consist of rock outcrop, elsewhere the drift cover supports a continuous tundra vegetation. Drainage is generally poor, and small lakes and swamps are abundant. The terrain becomes more rolling and the thickness of the drift cover increases towards the east.

1.2.1.5 Borden Upland

Borden Peninsula (Figure 7, Subdivision 5; NTS Maps 48C: Arctic Bay, 48D; Navy Board Inlet, 48B: Moffet Inlet, and 48A: Milne Inlet) is much more complex than Brodeur Peninsula. The central part of the area is a dissected plateau that reaches an altitude of about 1,220 m and supports local ice fields and small ice caps. Throughout much of the peninsula, traces of horizontal upland surfaces can be found at elevations of 360 to 600 m, but the deeply-incised valleys, and differences in structures and lithology, prevent correlation over large distances. The physiography of Borden Peninsula can be described in terms of five distinct units.

1: The hills and plateau present on northern Borden Peninsula are separated from Lancaster Sound by a lowland. This lowland is approximately 12 km wide, and its southern boundary begins at a point east of Cape Joy, extends eastward roughly parallel to the coast, and curves southward along

the western shore of Navy Board Inlet to a point located approximately at the latitude of Tay Bay. The area is underlain by sandstones and limestones overlain by rubble and felsenmeer. Along the coast, raised beaches form a strip that rarely exceeds 200 m in width. Inland from this strip, discontinuous bluffs and cliffs rise to about 100 m, and form a transition to an undulating surface dissected by valleys. This surface extends inland to the steep rise that forms the beginning of the inland plateau. Rivers cross this lowland, at first in deeply-cut valleys, then in less entrenched ones until they reach the coastal bluffs where they are deeply incised again. They generally reach the coast through ravines or wide gaps in the bluffs. A few wide sand and gravel deltas occur at the mouths of some rivers, and are presently being cut by their own streams. Drainage is in general, good, and there are no lakes. Between Cape Joy and Cape Charles Yorke, the shoreline consists of a gently sloping, unconsolidated backshore fringed by a continuous gravel and sand beach. A similar shoreline type extends about 20 km eastward from Cape Charles Yorke. East of this point to Bluff Head, sections of steep rocky backshores devoid of sediment alternate with steep, unconsolidated backshores fringed by continuous gravel and sometimes sand beaches. Southeast of Bluff Head to a point opposite Tay Bay, sections of steep, moderately sloping, and gently sloping, unconsolidated backshores fringed by continuous gravel, and sand beaches alternate with steep rocky backshores either devoid of sediments or with pocket beaches of gravel and sand.

2: A distinct physiographic unit extends south of the northern coastal lowland and covers a large part of Borden Peninsula. It is bounded to the south by a line that joins the head of Strathcona Sound, to the valley of Alfa River where it joins Tremblay Sound. The area consists of a complex plateau subdivided by large valleys into many small interfluves. Major valleys combine and form a narrow, hilly lowland that crosses along the middle of this area from Elwin Inlet to the south of Navy Board Inlet. North of this lowland, the plateau was once continuous at an average elevation of about 1,000 m, and is underlain by dipping and gently folded sedimentary rocks and a few restricted areas of granitic rocks. Erosion of this plateau has broken it up into smaller units separated by deep and steep-sided valleys. South of the northern coastal lowland, the area is rugged but relatively low. Hill tops and valley sides are covered with rubble, and the same sediments are present but reworked on valley floors. In this northern part, interfluves are flat and cover large areas. Further south, the plateau sub-units are higher and larger, and support small ice caps. There are six major ice fields. The largest is dome-shaped, and covers about 250 km^2 east of Elwin Inlet. In the east, two of the main ice fields cover most of the Archean rocks that outcrop west of Low Point. Four outlet glaciers reach Navy Board Inlet from the easternmost ice field.

South of the narrow lowland to the southern limit of the plateau, the area consists of a second upland terrain that resembles its northern counterpart in many respects. The area is underlain by sedimentary rocks and the plateau surface has a remarkable uniformity, with an elevation that varies from 460 m near Strathcona Sound to 550 m further to the east. The plateau terminates on its southeastern side in a gentle, stepped, and

probably structurally-controlled slope along which runs the valley of Alfa River. Although uniform, the plateau is deeply cut by valleys that range from youthful tributary ravines to mature, wide, U-shaped valleys in which the river meanders. Dissection of the plateau is most pronounced between Strathcona Sound and Elwin Inlet. In this southern part, there are no lakes and no ice fields, except for a few very small ones (the largest covering about 18 km^2). Near the summit of interfluves, and at the source of many rivers, small snow patches persist from year to year.

The shores of Strathcona Sound consist of steep and moderately sloping, rocky backshores with continuous gravel, sand, and boulder beaches that alternate with sections of gently sloping, unconsolidated backshores fringed by continuous gravel and sand beaches. This last type of shoreline occurs most frequently where tributary valleys join the sound. Northeast of Strathcona Sound, to the first major delta 15 km northeast of the mouth of Elwin Inlet, the shores adjacent to Admiralty Inlet, as well as those of Baillarge Bay and Elwin Inlet, have steep or moderately sloping, rocky backshores with gravel, sand, and boulder beaches. These are occasionally interrupted by relatively short sections of shorelines consisting of gently or moderately sloping, unconsolidated backshores fringed by continuous gravel beaches. From the first major delta northeast of Elwin Inlet to Cape Joy, shorelines have unconsolidated backshores of variable slope, and are generally fringed by continuous sand, or sometimes gravel, beaches.

On the eastern side of the peninsula, south of a point opposite Tay Bay, the shoreline consists of sections of gently and moderately sloping, unconsolidated backshores fringed by continuous gravel and sand beaches, occasionally interrupted by sections of steep, rocky backshore either devoid of sediments or with continuous gravel and sand beaches. Southeast of a point located opposite Canada Point to Low Point, the shoreline takes on a different character. Backshores are generally steep and rocky, and long sections of coastline have essentially no beach. These sections alternate with four relatively narrow glacier tongues, of which three reach the inlet. When beaches are present, they are generally continuous and made up of boulders. Immediately south of Low Point, the slope of rocky backshores decreases and gives way to a shoreline that consists of gently or moderately sloping, unconsolidated backshores fringed by sand or gravel beaches. Considerable sections of the coast have wide intertidal zones of sand. This type of coastline extends south to Alfred Point. Southwest of this point, the shores of Tremblay Sound are rocky and steep, except in front of major tributary valleys. There, the steep, rocky backshores and gravel or boulder beaches give way to moderately or gently sloping, unconsolidated backshores and continuous gravel beaches.

3: The central plateau area is bordered on the south by a relatively narrow (16 to 32 km-wide) belt of land that crosses Borden Peninsula in a northwest to southeast direction, from Strathcona and Adams sounds to Milne Inlet. This distinct belt of land is bounded to the north by the valley of Alfa River and its continuation to the northwest. Its boundary to the south is much less distinct but, from east to west, it runs north of Uvajo

Mountain, and follows the tributary of the Robertson River that runs north of the Magda Plateau and its continuation to Adams Sound. Within these limits, the land is characterized by a series of gabbro dikes that strike northwest to southeast, and by a plateau heavily dissected by broad and steep-walled valleys. The dikes themselves modify the topography considerably. They range in width from a few metres to 120 m, and some stand out vertically up to 60 m above the surrounding sedimentary rocks. The sedimentary rocks have cliffs and bluffs that also strike northwest to southeast. Most cliffs have talus of colluvium and some areas are overlain by a thin cover of ground moraine.

On the western side of Borden Peninsula, the southern shoreline of Strathcona Sound and those of Adams Sound are characterized by high cliffs. Backshores are generally steep and rocky and fringed by continuous gravel or sand beaches. There are a few exceptions to this: the northern and southern tips of Uluksan Peninsula, and junctions of major tributary valleys with the sound, are characterized by gently sloping, unconsolidated backshores and continuous beaches of sand and gravel. The southern coast of Uluksan Peninsula, the St. Georges Society Cliffs, are different too in that they are mostly devoid of sediment, except for a few pocket beaches of sand and gravel along the southwestern part and immediately east of Arctic Bay.

Along the southeastern part of Borden Peninsula, the shorelines at the head of Tremblay Sound, and those along western Milne Inlet are generally steep and rocky. Beaches are generally continuous and made up of coarse sediments such as boulders and gravel. Where major tributary valleys reach Tremblay Sound or Milne Inlet (the valleys of Alfa, Beta, and Gamma rivers, and the northern shore of Fairweather Bay) shorelines consist of gently sloping, unconsolidated backshores and gravel beaches.

4: On the western side of Borden Peninsula, between Adams Sound and the entrance to Moffet Inlet, there is an area of about 3,840 km^2 that is characterized by a very dissected upland surface. In this area, the sedimentary rocks form blocks that are tilted with a gentle slope to the south and a steep one to the north. Rivers are generally deeply incised, and cliffs are present along the coast and along the lower part of river valleys. Inland, hills reach elevations of over 600 m, and locally, especially to the southwest, relief can exceed 100 m. There are a few small ice fields, and most appear to be surrounded by a barren rim with many semi-permanent snow patches. The absence of vegetation may indicate that these ice fields were larger during the recent past (Beschel, 1961). Glacial sediments appear to be restricted to some terminal moraines in the lower parts of valleys near Admiralty Inlet. Most river valleys are floored with deposits of sand and gravel.

The southern shore of Adams Sound and those of Fleming Inlet and Fabricius Fiord are steep and rocky, and are generally fringed by a narrow, continuous beach of gravel or sand. Where major valleys reach Adams Sound, backshores are gently sloping, unconsolidated and fringed by a continuous sand beach. The eastern shore of Admiralty Inlet, between Adams Sound and the valley of Eqalulik River, consists of moderately sloping, rocky

backshores and continuous gravel and sand beaches. South of the valley of Eqalulik River, around Levasseur Inlet, Stephens Headland, Red Valley to the base of the Nauyat Cliff, the shoreline consists of a gently sloping backshore fringed by a continuous gravel beach.

East of this dissected upland area, the terrain consists of a series of hill ranges separated by drift-covered lowlands and plateaus. The terrain is poorly drained, and numerous swamps and marshes are present to the west of Magda Plateau. The area may have been filled, during the recent past, by a proglacial lake. In the western part of this area, the terrain is smooth and rolling, with a local relief of about 30 m, and generally overlain by a thick cover of unvegetated drift. To the east, a well-defined block stands out steeply northeast of Magda Plateau between the valleys of Robertson River and its northern tributary. The summit of this area is occupied by a 100 km^2 ice cap that reaches an elevation of about 960 m. On the northern side of the ice cap, an outlet glacier descends into the valley tributary to the Robertson River.

Further to the east, there is at the head of Tremblay Sound and Milne Inlet, a belt of mountains, the Uvajo and Tununek mountains. These reach altitudes of over 500 m, sometimes within very short distances of the sea (Bruce Head) and are underlain by Archean igneous and metamorphic rocks. The valley west of Koluktoo Bay extends as a broad lowland to the interior part of Robertson River. This valley is three to five km wide and is filled with silt and clay deposits through which bedrock outcrops occasionally.

The southwestern part of Milne Inlet has steep, rocky shorelines, except along the western extremity of Koluktoo Bay. There the shoreline consists of a gently sloping, unconsolidated backshore fringed by a continuous sand and gravel beach. A submerged delta is present at the northwestern part of Koluktoo Bay and the shoreline includes an exceptionally wide intertidal zone of sand.

5: The south central part of Borden Peninsula, between Admiralty and Milne inlets, is occupied by a dissected plateau underlain by sedimentary rocks. The plateau surface has an average elevation of about 300 m and is dissected in its central part by deep valleys. The slope of the sides of valleys becomes steeper towards the perimeter of the plateau. Near the margin, the plateau is subdivided into a series of blocks by vertically-walled gorges that are over 150 m deep. The upper surface of these blocks and the central part of the plateau are overlain by rock fragments that show no sign of transport. There is very little vegetation. Most gorges have flat bottoms consisting of gravel deposits. The wider valleys have deposits of sand, and support a tundra vegetation. The southern part of this unit follows an east-west trending scarp that denotes the transition from the sedimentary cover to igneous and metamorphic rocks.

To the east, elevation decreases progressively, and the plateau area is replaced by a series of low hills south of Milne Inlet. In this area, the substrate consists of sedimentary rocks, limestones covered by a bouldery drift. Locally, the till shows drumlins oriented in a north-south direction. In the south, Phillips Creek forms a wide, flat-bottomed valley that follows the contact between Paleozoic sedimentary rocks and the underlying Archean igneous and metamorphic rocks. The floor of the valley is filled with outwash and silt deposits into which the river has cut terraces.

1.2.1.6 Melville Upland

The Borden Upland is separated from the Foxe Basin Lowlands by a relatively narrow, triangular subdivision: the Melville Upland (Figure 7, Subdivision 6; NTS Map 47H: Phillips Creek). This area is characterized by a drift-covered, gently undulating upland surface lying at a mean elevation of about 240 m. The area is thought to be completely underlain by granitic rocks, even though bedrock outcrops rarely. A substantial thickness of till composed of granitic boulders in a sandy matrix is present throughout the area. This upland surface is uniform over large distances and its uniformity is interrupted only by a north-facing bluff, locally 30 m high, that extends along 160 km in the southeastern part. The plateau is also crossed by two deep valleys. The western one has been blocked by drift and forms a fiord and two lakes. The second valley is occupied by Gifford River. In its upper part, Gifford River is gently meandering and flows in a sandy bed between sand and boulder flats. Further along its course, it flows through a series of rapids and high, sandy, vegetated banks.

1.2.2 Foxe Plain

The western central part of Baffin Island is occupied by very large lowlands which, together, form the second major subdivision of the Arctic Lowlands (Figure 1; Figure 7, Subdivision 7). These lowlands have developed partly on Paleozoic sedimentary strata (mainly limestone) and partly on Precambrian Shield rocks. To simplify their description, they are subdivided into three major units: the area north and west of Steensby Inlet, the islands of eastern Foxe Basin and Baird Peninsula, and the area north and west of Nettilling Lake.

1.2.2.1 Area North and West of Steensby Inlet

South of the Melville Upland, the granitic rocks are overlain by horizontally-bedded limestone and form a distinct rectangular area that covers more than 2900 km^2 (NTS Maps 47H: Phillips Creek; 37G: Icebound Lakes; and 37F: Steensby Inlet). To the west, the sedimentary strata have been dissected into broad ridges separated by flat terrain. These flat areas reach an elevation of 240 m and are generally overlain by a thick

cover of drift through which granitic rocks sometimes protrude. Hillsides have steps because of the horizontal stratification of the limestone, but these steps are largely obscured by drift. In general, drainage is well developed, except in the northern part where bogs and lakes do occur.

To the east, the drift cover is broken by numerous granitic knobs and ridges that increase the local relief to about 60 m. Drainage is to the west, towards the large lake located east of Inuktorfik Lake. Numerous shallow lakes and bogs occur along rivers and streams.

The eastern side of Steensby Inlet is marked by a 300 m-high scarp. West of this scarp, the terrain is flat and dips gently towards the south. The plain surface is occasionally broken by steep-sided knobs that reach 20 m in height. The area is underlain by gneiss, but a thick cover of drift, with occasional large granitic boulders, is present everywhere. Drainage is poorly developed, and shallow lakes of irregular size and shape are abundant.

West of Steensby Inlet, Thalbitzer Peninsula rises gently to an elevation of about 100 m. Fragments of sandstone and limestone, together with occasional large granitic erratics, form the drift that covers most of the surface of the peninsula. Rock outcrops are rare, and drainage is so poorly developed that streams and lakes are largely absent. The few that are present are invariably shallow. In the centre of the peninsula, the surficial sediments show a faint ridging and may consist of a washed till. Vegetation is either very sparse or absent, and raised beaches occur to a considerable distance inland.

1.2.2.2 Foxe Basin Islands and Baird Peninsula

The low islands present in eastern Foxe Basin (NTS Maps 37C: Koch Island, 37B: Spicer Island, 37A: Foley Island, 36N: Prince Charles Island, 35O: Air Force Island, and 36P: Hantzsch Bay) are composed mainly of Paleozoic limestones, and it seems that they had already been reduced to level plains before the Pleistocene. During the Pleistocene, ice moving southwestward left considerable deposits of till, and most islands were subsequently submerged either completely or to a large extent. Raised beaches can now be found at all elevations within the areas that have been submerged. As reworking of sediments requires wave action, raised beaches are usually better developed on steeper sections than on a flat terrain. In fact, the orientation of the till present in low, flat areas is still visible at several locations, even though it has undergone an identical submergence history.

On many islands, beaches are devoid of vegetation, but marshes are often present between successive raised beaches near present sea level. Numerous lakes are present on the till surface; most of these drain directly through their substrate rather than by a drainage system. Many lakes are fringed by marshes.

Jens Munk Island is separated from Thalbitzer Peninsula by a shallow, six km-wide strait. The eastern two-thirds of the island is underlain by limestone strata and forms a plateau that rises to 150 m. Most of the plateau is surrounded by a band of raised beaches, and a few low areas have a thick cover of alluvial debris and numerous lakes. The western third is underlain by Precambrian rocks and has a low but rugged surface. Large parts of the coast are associated with wide intertidal zones and extensive shoals.

Koch Island rises in its central part to a low plateau that reaches an elevation of about 25 m. Raised beaches occur almost to the centre of the island. The nearshore zones are all very shallow and the two bays on the northeastern side of the island are dry at low tides.

Rowley Island, although low (maximum elevation of about 60 m), is composed of a series of barren plateaus. Rock outcrops are rare, and in general, the surface is composed of rock fragments and reworked glacial deposits. Numerous lakes are present and the island is bordered by a band of raised beaches. Most bays located on the northern part of the island, except in the northwestern part, are fringed by wide intertidal flats.

Bray Island consists in a monotonous plain of drift with numerous lakes and swamps. It is fringed by limestone gravel beaches and very wide (three km) boulder intertidal flats.

Baird Peninsula projects southwestward into Foxe Basin. It is approximately 65 km long and 40 km wide. Most of the peninsula consists of a low, very poorly drained plain with numerous lakes and swamps. At its southwestern extremity, a low limestone plateau reaches an elevation of about 30 m. Along this southwestern part, limestone strata outcrop and form blunt headlands. Elsewhere, the coast is fringed by very wide intertidal zones.

The Spicer Islands are all small and very low. Their surface consists of limestone fragments reworked into raised beaches. They all have very wide intertidal zones, especially North Spicer Island.

Foley Island is approximately 50 km long and 22 km wide. In its northeastern part, Precambrian gneisses outcrop below Anderson Bluff and are overlain by limestone. This limestone forms sharp east and north-facing scarps that rise to an elevation of about 80 m. Inland from these scarps, the surface slopes gently westward and forms a very wide (four km) intertidal flat.

Prince Charles Island is the largest of the islands of eastern Foxe Basin. It consists of a broad plain of very low relief that reaches a maximum elevation of about 60 m (Fraser, 1953). A low plateau extends north to south on the western side of the island, and is fringed by numerous limestone raised beaches. The eastern part of the island is lower and less well drained.

Air Force Island has a core of gneissic rocks that form low ridges extending south from Fee Peninsula to the centre of the island. Offshore Fee Peninsula, the nearshore zones are the only parts of the island where water is relatively deep. The rest of the island has a sedge and grass tundra vegetation, numerous shallow ponds, and a very poorly developed drainage.

North and South Tweedsmuir islands have a barren plateau surface surrounded by a lower, drift-covered terrain. On South Tweedsmuir Island, raised beaches occur inland to an elevation of 20 m. There are relatively few lakes, and shores consist of either bedrock devoid of sediment or bedrock fringed by gravel beaches. Nearshore zones are generally very shallow.

1.2.2.3 Area North and West of Nettilling Lake

Between Wordie and Piling bays, a belt of upland forms the western coast of Baffin Island (NTS maps 37A: Foley Island, 36P: Hantzsch Bay, 26M: Hantzsch River, 36I & J: Koukdjuak River, and 26L: Nettilling lake). The upland surface has an elevation of 100 to 200 m and is underlain by a closely-jointed granitic gneiss. To the east, the area is bordered by steep bluffs, and includes lakes that often lie 60 m below the general level in joint-controlled valleys. Other valleys are floored with silt and marshes. At several locations, the coast consists of low bluffs; elsewhere, small, gently sloping silt plains fringed by wide (six km) intertidal flats of mud and boulders make up the shoreline.

East of this belt, an area of lowland with summit elevations at 240 to 270 m rises gently east to the Baffin Uplands and north to the Gillian Hills. The area is composed of low ridges separated by valleys floored with glacial drift, numerous lakes, and occasionally, eskers and sand or gravel terraces. Towards the head of Straits Bay, the structural control changes from a linear to a folded one, and the bedrock ridges are generally curved. Swamps occur frequently in coastal valleys. Silt is being deposited at the mouths of many rivers, and forms very shallow nearshore zones, especially at the head of Piling Bay. There are also extensive areas that are very flat and swampy. Elsewhere, the coast is either fringed by a wide (two km) intertidal zone of mud and boulders, or consists of low bedrock bluffs with few pocket beaches or no sediment at all.

West of the Central Baffin Hills and the Nettilling Upland, the area is low, gently undulating with small ridges that become progressively more rugged eastward. Numerous preglacial valleys have been filled with drift and are now occupied by a vast network of small lakes. Part of the lowland is drowned in the interior and forms Amadjuak and Nettilling lakes. Near these lakes, the drift cover is thinner than anywhere else. Low bedrock outcrops form numerous small islands in the eastern part of Nettilling Lake. On the southern and southeastern sides of this same lake, the drift cover is thicker and is composed of boulders in a clay matrix and a few sandy eskers. Along the coast, the sediments have been reworked into an

extensive system of raised strandlines. West of Nettilling Lake, a level plain appears to have emerged from the sea only very recently (Manning, 1943). The plain consists of marine clay and peat, with some limestone outcrops at the bottom of large streams. These surficial deposits average two to three m in thickness (Soper, 1930) and are covered by a tundra vegetation of grass and sedge. Numerous shallow, circular lakes are present, and show various stages of infilling by vegetation. Along the coast, very wide intertidal mud flats, three to six km wide, fringe a highwater line that is often the site of a line of boulders. Nettilling Lake drains across this plain by the Koukdjuak River. This river appears to be in a youthful stage; it follows a very straight course, it is very wide (two to six km) and shallow (one to three m).

1.3 Coastlines of the Eastern Arctic

The types of coastlines present in the eastern Arctic have been described either qualitatively at a regional scale (Canadian Hydrographic Service, 1970; Dunbar and Greenaway, 1956; McGill University, 1963a, b and c; Miller et al., 1980; Owens, 1977; and Stevenson, 1978), or into greater detail as a result of site-specific studies (Barrie et al., 1978; Gilbert, 1982, 1980, 1978; and Taylor, 1980a).

In 1978, a coastal mapping program was undertaken by Petro-Canada. It began by the interpretation of vertical air photos and later included two summers of field work done in cooperation with personnel from the Geological Survey of Canada and Bedford Institute of Oceanography (McLaren et al., 1981). The area studied during this program extended from southern Ellesmere Island in the north to the northwestern part of Brodeur Peninsula in the west and Cape Dyer in the south.

Coastlines were described and mapped according to a system developed at the Geological Survey of Canada (McLaren, 1980). This system describes coastal zones by using any combination of the following symbols:

1) s, m, or l to denote the slope of backshores (s = steep, 45° to 90°, m = moderately sloping, 15° to 45°, and l = gently sloping, 0° to 15°);

2) R or U to denote the nature of backshore (R = bedrock, and U = unconsolidated);

3) (c) to denote the presence of colluvial debris;

4) :, -, or / to indicate the amount of beach present in a given coastal map unit (: = more than 80 per cent, - = 20 to 80 per cent, and / = less than 20 per cent);

5) c, p, or b to indicate the type of beach present in a coastal map unit (c = continuous, p = pocket, and b = barrier);

6) the letter B for beach; and

7) b, g, s, or $ to denote the dominant beach sediment type in a given coastal unit (b = boulders, g = gravel, s = sand, and $ = silt).

This mapping system, like most other ones, looks terribly confusing. In reality, it is probably less so than many other ones as the mnemonic nature of the components helps in 'reading' the description of coastal units. For example, sR signifies a steep (s) rocky (R) backshore without beach; sRicBb signifies a steep (s) rocky (R) backshore with colluvial debris ((c)) and more than 80 per cent (:) of continuous (c) beach (B) made up of boulders (b); and lU-bBs denotes a gently sloping, unconsolidated backshore with 20 to 80 per cent of barrier beach made up of sand. The results consist of a series of 1:125,000 maps and these maps were subsequently analysed by Sempels, 1982.

1.3.1 General Coastal Characteristics

At this point, it is important to remember that the following discussion describes the shorelines of eastern Devon Island, northern Brodeur Peninsula, Borden Peninsula, Bylot Island, and northern and eastern Baffin Island. The shorelines of eastern Foxe Basin are not included in this study. It is obvious from the physiographic description that the inclusion of these shorelines would alter results considerably.

The study area, as defined above, comprises 11,700 km of coastlines and, according to this mapping system, 94 distinct types of shorelines (Table 1). Approximately 75 per cent of all coastlines include a beach and less than one per cent consists of glacial ice. Backshore slopes are predominantly steep (~60 per cent), and least frequently, gently sloping (~14 per cent). Backshores consist mainly of bedrock alone (~60 per cent), or bedrock with colluvial debris (~20 per cent). Unconsolidated backshores occur in less than 25 per cent of shorelines. Beaches, when present, are mainly continuous (~65 per cent) and made up of boulders (~35 per cent) or gravel (~28 per cent).

From these overall statistics, one may get the impression that the most common type of shoreline in the eastern Arctic consists of a steep rocky backshore fringed by a continuous boulder beach. Although not far from the truth, it is not the case. This difference illustrates well the fact that it is essential to understand what one is doing when dealing with regional compilations. The above statistics (Table 1) result from the compilation of coastal characteristics taken independently. The description of coastal segments requires that these characteristics be considered together as they occur in reality.

In the eastern Arctic, the five predominant types of coastal zones, ranked in order of decreasing abundance, are:

1) a steep, rocky backshore without beach (sR,22.5 per cent or 2634 km);

Length of Coastline (km)		11708
Number of Map Units		2569
Number of Coastal Zones		94
Coastline:	Beach Absent	26.1%
	Beach Present	73.3%
	Glacial Ice	0.6%
Backshore Slope:	Steep	60.7%
	Moderate	25.2%
	Low	13.5%
Backshore Material:	Bedrock	56.3%
	Bedrock with Colluvium	20.8%
	Unconsolidated	22.2%
	Unconsolidated with Colluvium	0.1%
Amounts of Beach in Map Unit:	More than 80%	55.5%
	20 - 80%	9.5%
	Less than 20%	8.3%
Types of Beach:	Continuous	64.5%
	Pocket	8.1%
	Barrier	0.7%
Dominant Beach Sediment:	Boulder	34.5%
	Gravel	28.2%
	Sand	10.3%
	Silt	0.3%

Table 1 General coastal characteristics of the eastern Arctic.

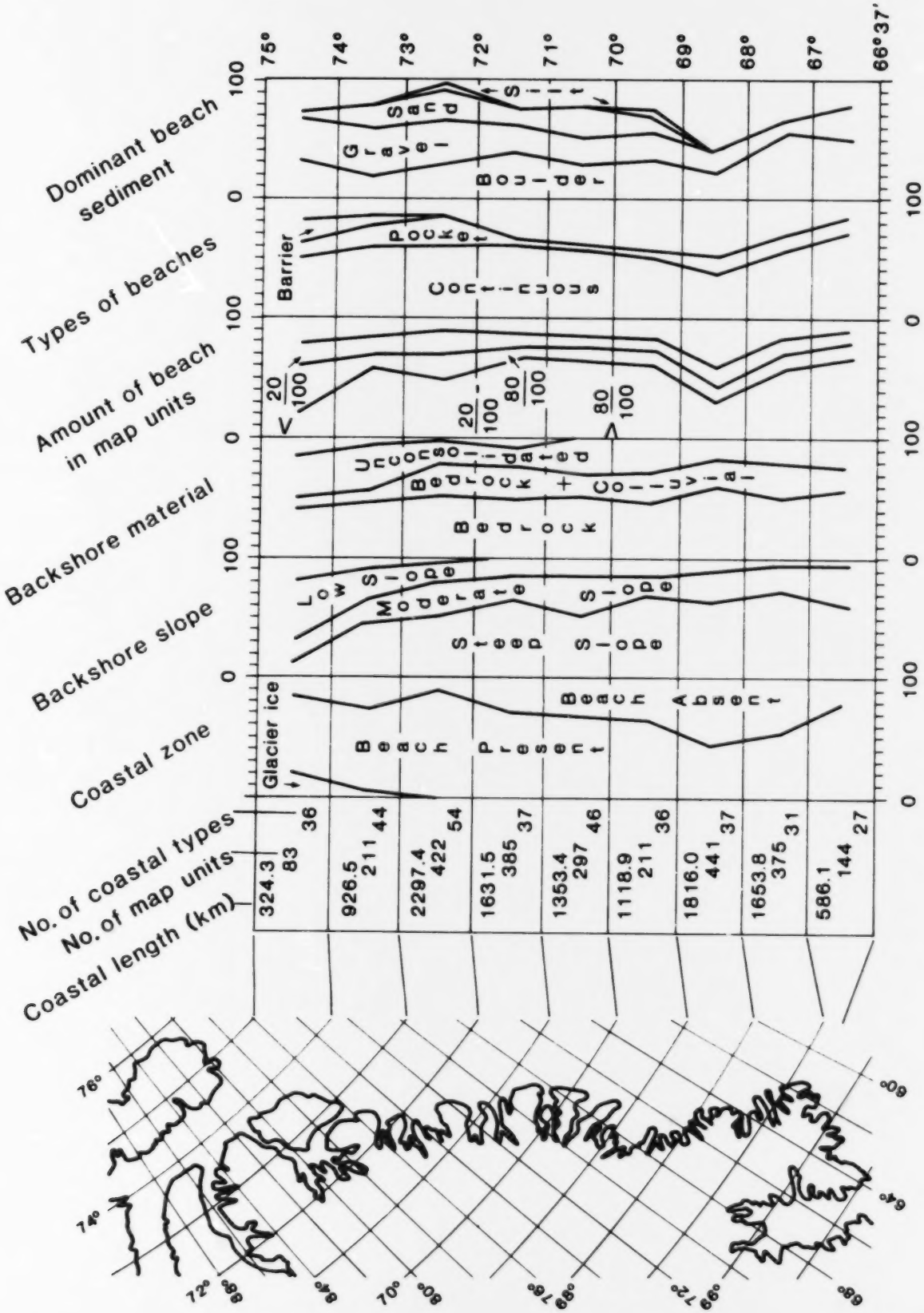


Figure 8 Variations of coastal characteristics according to latitude, after Sempels, 1982.

2) a steep, rocky backshore with colluvial debris and a continuous boulder beach (sR(c): cBb, 18.4 per cent or 2154 km);

3) a steep, rocky backshore without colluvium but with a continuous boulder beach (sR:cBb, 5.4 per cent or 632 km);

4) a moderately sloping, unconsolidated backshore with a continuous gravel beach (mU:cBg, 5.2 per cent or 608 km); and

5) a gently sloping, unconsolidated backshore with a continuous gravel beach (lU:cBg, 4.6 per cent or 538 km).

Together, these five most abundant types of shorelines make up 56.1 per cent or 6568 km of all coastlines present in the eastern Arctic. The remaining 43.9 per cent or 5140 km comprise 89 other types of shorelines.

Results of compilations of coastal characteristics and coastal types can appear contradictory, and it is important to understand why this is so. For example, one important characteristic is that coastlines are often associated with a beach. Yet the most abundant type of coastline consists of a steep, rocky backshore without beach. Both facts are true, and the contradiction disappears when one realizes that this type of coastline, a steep rocky backshore without beach - although being the most abundant type, makes up only 22.5 per cent of all coastlines. The remaining 77.5 per cent include 93 other types of coastlines, most of which include a beach.

1.3.2 Variations of Coastal Characteristics According to Latitude

A knowledge of overall coastal characteristics and relative abundance of shoreline types is useful in understanding the geological environments of the eastern Arctic; however, these parameters fail to convey a measure of their variability in space. If a steep, rocky shoreline without beach is the most abundant type of coastal zone in the eastern Arctic, is one likely to encounter it as frequently on southeastern Devon Island as in the area of Home Bay? The answer is, as one might expect, no.

Variations of characteristics within the overall study area can be detected by grouping coastal data according to increments of one degree of latitude (a purely arbitrary choice), and by comparing graphically the characteristics of each group (Figure 8). Within the study area, glacial ice is an important type of coastline in the northernmost part. North of 73°N, backshores are generally gently sloping and are often made up of unconsolidated sediments; south of 73°N, they are mostly steep and made up of bedrock. Although beaches are often present throughout the entire area, they occur most frequently between 72° and 73°N and least frequently between 68° and 69°N. Continuous beaches are present everywhere; pocket beaches (small beaches bounded by rocky headlands) occur more frequently north of 72°N and south of 69°N; and barrier beaches (elongated, narrow beaches enclosing a lagoon) occur more frequently north of 73°N. Coarse beach sediments (boulder and gravel) occur throughout the area, and finer ones (sand and silt) occur mainly in beaches located between 69° and 73°N.

It is also possible to detect variations in types of coastlines by a similar grouping of data, and by comparing the dominant types of each group (Figure 9). On southeastern Devon Island, between Croker Bay and Philpots Island (Figure 9, Line 1), the five most abundant types of coastlines are:

1) a gently sloping, unconsolidated backshore with a continuous gravel beach (lU:cBg, 52 km);

2) glacial ice (Gl, 40 km);

3) a steep, rocky backshore without beach (sR, 19 km);

4) a moderately sloping, unconsolidated backshore with a continuous gravel beach (mU:cBg, 18 km); and

5) a steep, rocky backshore with colluvial debris and a continuous boulder beach (sR(c):cBb, 15 km).

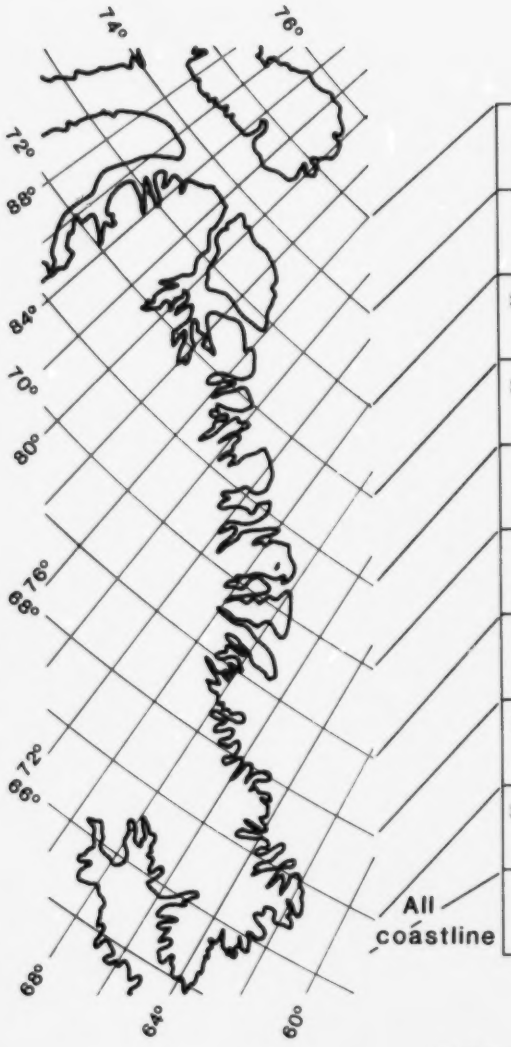
Together, these five types occupy 43.7 per cent (142 km) of all shorelines of the area and the remaining 56.3 per cent (182 km) comprises 31 other types of coastlines.

South of 74°N to Cape Dyer (66° 37'N), the dominant type of coastline becomes a steep, rocky backshore, with or without colluvium and boulder beach. In Home Bay, between 68° and 69°N, the dominant type of coastline is a steep, rocky backshore without beach and it makes up, by itself, 45.6 per cent of all coastlines of that area.

The combined abundance of the five predominant types of coastlines (Figure 9, Column 6) is a measure of the importance of these five types in the area under consideration. For example, the five most abundant types of coastlines make up only 39.1 per cent of all shorelines located between 72° and 73°N. In the area of Home Bay, the five dominant types make up 78.1 per cent of all coastlines. The overall diversity of coastal types in a given area is indicated by the number of remaining types (Figure 9, Column 7). In the eastern Arctic, the diversity of coastlines is greatest between 70° and 71°N and least between 67°N and Cape Dyer.

1.3.3 Coastal Processes

After understanding the coastal characteristics and coastal types of the eastern Arctic and how these vary in space, another aspect remains to be considered; that is, how these characteristics vary with time. Broadly speaking, coastal processes can be defined as any natural phenomena that induce a change in the coastal environment. For example, erosion or deposition are natural phenomena that modify the characteristics of coastlines. A geographic description of coastal processes that are taking



Types of Coastal Zones

					Combined abundance	Number of remaining types of coastal zones
IU:cBg 15.7%	GI 12.2%	sR 5.8%	mU:cBg 5.5%	sR(c):cBb 4.5%	43.7%	31
sR 16.8%	IU:cBg 16.0%	sR:cBb 7.7%	sR:cBg 6.6%	sR(c):cBg 5.6%	52.7%	39
sR(c):cBb 10.3%	sR:cBb 9.6%	sR 7.0%	sR-cBb 6.5%	mR:cBg 5.7%	39.1%	49
sR(c):cBb 27.5%	sR 23.5%	sR:cBg 8.0%	IU:cBs 8.0%	sR:cBb 6.6%	73.6%	32
sR 25.3%	sR(c):cBb 21.7%	mU:cBg 11.5%	IU:cBs 10.4%	mR:cBg 3.6%	72.5%	41
sR 31.0%	sR(c):cBb 27.5%	mU:cBg 10.3%	IU:cBg 4.5%	mR:cBg 2.7%	76.0%	31
sR 45.6%	sR(c):cBb 13.9%	mR/pBg 7.4%	mR 7.2%	mU:cBg 4.0%	78.1%	32
sR(c):cBb 27.2%	sR 21.1%	mU:cBb 9.1%	sR:cBb 6.8%	mR:cBb 5.9%	70.1%	26
sR(c):cBb 16.7%	sR:cBb 14.1%	sR 8.3%	mR:cBb 8.0%	mU:cBg 5.9%	53.0%	22
sR 22.5%	sR(c):cBb 18.4%	sR:cBb 5.4%	mU:cBg 5.2%	IU:cBg 4.6%	56.1%	89

All coastline

75° 74° 73° 72° 71° 70° 69° 68° 67° 66°37'

Figure 9 Variations of types of predominant coastal zones in the eastern Arctic, (see text for explanation of symbols) after Sempels, 1982.

place in an area as large as the eastern Arctic would require endless field studies as well as an endless report to contain it. It is nevertheless possible to describe the parameters that influence and give rise to coastal processes. Of these parameters, the major ones are winds, waves, tides, and ice.

1.3.3.1 Winds

Winds play an important role in modifying coastlines. They do so mainly by the generation of waves and by the direct transport of fine sediments. During periods of strong winds, large volumes of sand can be transported inland or offshore from a beach. Sand can also be deposited on a beach from an inland source, as for example, a dune.

Strong onshore winds can also cause a rise in water levels at the shoreline by literally piling up water against a coast. This phenomenon is called a storm surge, and can happen on any coast, especially those with large embayments. Storm surges can modify a beach by subjecting a higher part of the shoreline to wave action. As a result, storm surges can have a considerable effect on the redistribution of backshore sediments or on backshore erosion.

In the eastern Arctic, the prevailing winds are out of the west or northwest. During summer months, winds are often out of the southwest due to the influence of a large low pressure system centred over Iceland and Greenland. In general, there are little differences in average wind velocities between the two seasons. When compared to more southerly latitudes, wind velocities are generally lower. Topography can modify wind patterns to produce strong local winds.

1.3.3.2 Waves

Waves are generated by the interaction of wind and water surface. The energy of waves is determined partly by wind velocity and duration, and partly by fetch. Fetch is the length of open water over which the wind can blow. This distance limits the time during which energy can be transferred from wind to waves (Komar, 1976) and consequently is a parameter that limits the maximum energy level at the shoreline (King, 1972). With respect to fetch, it is interesting to note that in the eastern Arctic, 71.9 per cent (8416 km) of all coastlines are exposed to a fetch of less than 10 km; 12.5 per cent (1464 km) to a fetch of between 10 and 100 km; and 15.6 per cent (1828 km) to a fetch of more than 100 km (Figure 10).

The interaction of waves and a shoreline is a complex matter that has been, and still is, the subject of much research (Komar, 1976; King, 1972). The most important aspect of waves is that they dissipate energy in the littoral zone. In essence, when waves reach shallow water, they become progressively deformed due to bottom friction, to a point where they can

BAFFIN BAY
BAFFIN ISLAND
Fetch >100km
Fetch between 10 and 100km
Fetch <10km

become unstable and break. Two main types of waves can occur: sea or storm waves, found within the generating area, and swell waves generated by distant storms. Coasts open to swell and storm waves are high energy environments. The eastern Arctic is essentially a low energy environment, since 1) most of the coastlines are exposed to a limited fetch (Figure 10), and 2) waves are dampened by the presence of ice. It must be understood that in a low energy environment, levels of energy are low throughout the year but a section of coast can nevertheless be subjected to the occasional intense storm wave activity.

The energy transfer associated with the breaking of waves results in the redistribution of sediments. This redistribution can take place in the offshore, onshore or alongshore directions. During periods of high energy activity, the transfer of sediments often takes place in the offshore direction, that is, the beach is eroded. When this activity decreases, an onshore movement rebuilds the beach. In areas that have large seasonal differences in wave energy levels, these processes can lead to a summer-winter beach cycle. As described in the discussion on winds, this is not the case for the eastern Arctic. In this area, seasonal variations are less than those caused by storms, and in such a storm wave environment, storm-post storm cycles predominate.

When waves approach a beach at an angle other than 90°, the swash runs up the face of a beach at an oblique angle. The backwash runs back down in a direction that is determined by gravity alone, generally at 90° to the water line. The imbalance results in longshore or beach drift, and sediments that can be moved by incoming waves are transported alongshore. During the open water season, such movements occur frequently in the eastern Arctic.

When wave energy levels are very high or when the water level is high (during a storm surge or very high tide), wave activity can extend to higher parts of a beach and can result in overwash. During this process, sediments are thrown landward as waves break on a beach face. In beaches made up of coarse sediments, this can result in the build-up of a storm berm. In beaches made up of fine sediments, especially those backed by a low area or a lagoon, much material can be transported landward with the overwash. This process occurs in the eastern Arctic but its frequency is restricted by the length of open water season.

1.3.3.3 Tides

Tides result from the gravitational effects of the sun and moon on the water bodies of the earth. These effects have a magnitude that varies from one area to another. The relative effects of the sun and the moon also change regularly, and result in a monthly or seasonal variation in tidal range. High, or spring tides, occur twice a month and are highest twice a year at the spring and autumn equinoxes. Tidal changes are usually semi-diurnal cycles. A mixed diurnal and semi-diurnal cycle is often present, and causes irregularities in successive high and low tide levels. In the

eastern Arctic, the semi-diurnal tidal heights are largest in the northern end of Baffin Bay with amplitudes (half of the tidal range) of 80 cm, and decrease to 60 cm in Lancaster Sound. The amplitudes continue to decrease south to a zero value in the vicinity of an amphidromic point located off Cape Hewett. Further south, the amplitudes increase to about 80 cm off Cape Dyer and thereafter increase sharply further south. The diurnal tidal heights are more uniform throughout Baffin Bay, with amplitudes of 25 to 30 cm decreasing from north to south (Fissel, 1982).

Tides can affect coastlines in two ways: by the action of tidal currents, and by controlling the effectiveness of wave energy at the shoreline. Tidal currents result from the rise and fall of sea level. These currents are important in areas of high tidal range, or where the horizontal displacements of water are restricted by the distribution of coastlines. If horizontal currents due to tides are strong enough, they will affect the redistribution of sediments. The effectiveness of wave action is inversely proportional to the tidal range. This is because wave energy is dissipated over a larger vertical range in the case of a high tidal range (Hayes, 1975). Tides can also act as storm surges in making higher parts of beaches accessible to wave action.

1.3.3.4 Ice

In the eastern Arctic, ice may well be the most important parameter that affects and controls littoral processes. The most important effect of ice is to prevent wave generation and to protect the shoreline from wave action (Owens and McCann, 1970; and Taylor and McCann, 1976). Ice on water limits the fetch and dampens waves, and the immediate effect of ice is to reduce wave energy levels considerably. In the eastern Arctic, wave generation is limited to only a few months each year (McCann, 1973) and in some parts of the Arctic, some years may have no open water season at all.

Ice also forms on the shorelines (Short and Wisemann, 1973; 1974; McCann and Carlisle, 1972; Kovacs and Mellor, 1974; Adams, 1977) and acts as a barrier against waves. This zone of fast ice, often called an ice foot, forms before freeze-up and persists after break-up. Consequently, the length of time during which the shoreline is free of ice and subject to littoral processes is less than that of the open water period. It is generally thought that the presence of an ice foot acts as a protective cover on the beach until its melt, at which time sediment rafting can take place. Recent work at Cape Hatt indicates that the formation and growth of an ice foot may involve a redistribution of the substrate.

Ice can also affect beaches directly. When shorelines are free of ice, wind can act on the pack ice and push individual floes onto the beach. When the ice grounds, it can push sediments landward to form a ridge. Scouring also takes place in the nearshore zones. Grounded growlers can be rocked back and forth by waves, and can create a substantial depression while protecting the upper part of the beach. Upon melting, they often leave

large beach pits. The presence of ice in the intertidal zone also interferes with normal sediment transport (Taylor, 1981). Ice also occurs within sediments, as interstitial water freezes during periods of subzero temperatures. The presence of ice-bounded sediments in a beach can modify normal erosion rates (Taylor, 1980b).

1.4 Vegetation of the Eastern Arctic

Although arctic and alpine regions have unique climatic regimes, their vegetation is not unique and native species are few. All arctic vegetation types have counterparts elsewhere in the world. Many tundra species have a circumpolar distribution and can adapt to a wide range of environmental conditions and habitats. Arctic ecosystems are generally low in species diversity but high in numbers of individuals. The abundance and distribution of arctic flora is limited by environmental factors such as a short growing season, low temperatures, low precipitation, poor soil development or lack of soil nutrients, the presence of permafrost, and frequent winds. In general, vegetation is scarce, but locally favourable conditions can combine to create microhabitats with an abundant flora. In such favourable microhabitats, primary productivity may reach levels comparable to a temperate herbaceous meadow (Peterson, 1976).

Studies of arctic flora indicate that its distribution is governed largely by topography and substrate, and that regional vegetation types may not exist. Within relatively small areas, large differences in habitats can occur, and result in large differences in species composition and productivity. This is certainly the case of the fiord coast, mountains and plateaus of eastern Baffin and Devon islands, where the diversity of terrain is accomplished by a corresponding diversity in vegetation.

There is also a correlation between the nature of the substrate and the quantity and composition of plant cover. With the exception of locally luxuriant communities (Polunin, 1948), there is a consistently lower diversity and abundance of plants on the Paleozoic sedimentary strata of Brodeur and Borden peninsulas, and western Ellesmere, Devon, and Baffin islands.

Many different systems of classification of arctic vegetation have been used in the past. The relative abundance of plants has been recognized as high in the low Arctic, conspicuous and open in the middle Arctic, and very sparse in the high Arctic. Three main types of communities have been described in the Arctic Archipelago: tundra sedgelands and grasslands; polar semi-deserts; and polar deserts (Polunin, 1951). A classification of 1) fell-fields or rock deserts, 2) stoney lichen heaths, and 3) dwarf shrub lichen heaths has been used to describe the vegetation of Baffin Island. Wilson (1976) described the flora of Auyuittuq National Park in terms of fell-field communities, dwarf shrub-heath tundra, grassland tundra, shoreline communities, and snowpatches.

The absence of a recognized classification system makes spatial descriptions very difficult. It is nevertheless informative to consider some of the main types of communities.

Fell-fields or rock deserts occur where vegetation is scattered on upland surfaces, relatively bare lowlands, and sedimentary plains. Woody species and heath are lacking. Crustaceous and leafy lichens grow on rocks and barren flats, with other lichens, mosses, cushion plants and rosette species such as saxifrages, on surrounding stoney soils. Arctic poppy (Papaver radicatum), a hardy pioneer species, occurs along with tufted grasses and perennial herbs. Vascular plant cover is low, usually less than 20 per cent (Bliss, 1975). Fell-fields include both the polar deserts and the semi- deserts as defined by Polunin, and are found throughout the unglaciated uplands of Baffin and Devon islands.

Tundras differ from fell-fields, in having a continuous or nearly continuous cover of low shrubs, herbs, lichens, and mosses. Tundras occur on sheltered soil surfaces, and tundra meadows are normally found in coastal lowlands, fiord valleys, and some protected upland surfaces. A depauperated tundra, dominated by lichens, particularly Reindeer moss (Cladonia sp.), can be found in dry upland areas, on slopes, and on sandy tills. In these, small amounts of sedges, tufted grass, arctic willow (Salix arctica) and mountain avens (Dryas integrifolia) are often interspersed with the lichen cover.

Wet tundras have a denser vegetation cover, with grasses and rushes as predominant plants. In areas that are very wet, cottongrass (Eriophorun sp.) and damp sedges are abundant. On fine-grained soils where ground ice is present, frost heaving often produces a damp, tussock tundra. In favourable locations, as in the southern part of Baffin Island, a diverse vegetation of dwarf shrubs and woody species does grow.

There is a normal development of salt-tolerant grasses and sedges in shoreline areas of the Arctic. Communities are generally absent from foreshore zones because of wave and ice action. On dry sandy beaches above the high tide mark, an open vegetation dominated by sandworts (Arenaria peploides), Lyme grass (Elymus arenarius), and sea pinks (Armeria maritima) can be found. Where shores are made up of fine sediments such as sand or mud, a dwarfed salt marsh community can develop. This community is characterized by alkali grasses (Pucinellia officinalis), sedges, and scurvy grass (Cochlearia sp.).

Locally, luxuriant vegetation occurs throughout the eastern Arctic. Important in mountainous regions are snowpatch communities, which grow in areas where deep snow ensures a good supply of moisture during the summer. Depending on the persistence of the snow, vegetation can be very abundant. Snowpatch communities consist generally of herbaceous species, including sedge, saxifrage, mountain sorrel (Oxyria digyna), dwarf buttercup (Ranunculus pygmaeus), and some mosses. Vegetation can also be dense along stream banks, as these provide abundant moisture and shelter from the wind. Locally enriched microhabitats are found around bird nests, old Inuit dwellings, lemming burrows, and musk-ox meadows.

The spatial distribution of plants throughout the eastern Arctic is not well known. Site-specific botanical and ecological studies have been done on Bylot Island (Drury, 1962), in the vicinity of settlements on Baffin Island (Polunin, 1948), at Coburg Island (Muller, 1979), and at Truelove Inlet on northern Devon Island (Bliss, 1975), but no detailed map of vegetation has been compiled. The Truelove Lowland is a rich coastal plain, and the site of many years of ecological research under the International Biological Program. Although few in number and of small area, luxuriant tundra meadows such as this one can be very important to a large number of mammals and birds.

The abundance of microhabitats is a striking indication of the severity of the arctic environment. Every small advantage, whether protection from wind, better moisture, favourable exposure to sunlight, or nutrient enrichment of the soil, is exploited by plants.

2 Offshore Physiography of the Eastern Arctic

Early Arctic explorers made preliminary contributions to the understanding of the sea floor in northern regions. Their investigations were followed by hydrographic studies conducted by Russian oceanographers in the 1930's and 1940's, and in recent years by scientists from the Department of Mines and Technical Surveys, Bedford Institute of Oceanography, and the Geological Survey of Canada. Although the literature covering eastern Arctic offshore physiography may seem extensive, the data are derived mainly from isolated geophysical and geological surveys. As a result, theories regarding the origin and tectonic structure of the eastern Arctic are often conflicting. The area under consideration in the following description includes Smith Sound, Jones Sound, Lancaster Sound, Baffin Bay, and Davis Strait.

2.1 Smith Sound

Within Smith Sound, the regional elevation of the sea floor decreases from depths of 300 m below sea level in the north to 700 m below sea level in the south. Analyses of bathymetric contours lead Pelletier (1964) to suggest that the eastern side of the sound is formed by a drowned portion of the western coast of Greenland, and that no continental shelf is present. On the western side of the sound, a narrow continental shelf is possibly present off the southeastern coast of Ellesmere Island. The uniformity of depth along this shelf may be due to horizontal bedrock or a submerged erosional surface. As depths in the east are similar to those in the west, Pelletier (1964) concludes that submergence has occurred, and that topographic features have developed under subaerial conditions. In the northern part of Smith Sound, the relief of the sea floor shows an east-west trend. This alignment is more apparent in the east, where Pelletier (1964) has traced the extension of headlands and valleys from the land to the coastal portions of the sound.

The physiography of Smith Sound has been further defined by geophysical measurements (Keen and Barrett, 1973; and Hood and Bower, 1974; 1975). Seismic reflection and aeromagnetic surveys indicate the presence of two major geological trends. The first is a central bathymetric channel, 500 to 700 m deep, which follows the north-south Nares Strait lineament. South of the Thule area, this channel curves to the southeast. This change in direction may be due to glacial scouring in the vicinity of Inglefield Bay (Greenland). In the south, the acoustic and magnetic basement appears to be near the sea floor. To the north, smooth magnetic anomalies suggest that over four km of faulted and folded sediments may lie above the magnetic basement (Keen and Barrett, 1973). The second trend is a linear belt of negative anomalies running northwestward from Thule, between Parry and Carey Island (Greenland). Keen and Barrett (1973), and Hood and Bower (1974, 1975) suggest that these anomalies represent a sediment-filled, half-graben basin which reaches depths of 20 km in the east, and shallows to 10 km in the west. Although evidence is incomplete, Keen and Barrett (1973) suggest that minor faults and folds in the uppermost sediments have resulted from recent tectonic activity, as this sedimentary basin follows a trend similar to the Precambrian grabens in the Thule area. They also suggest that the basin defined in Smith Sound is of comparable age and origin.

2.2 Jones Sound

Jones Sound is approximately 550 m deep, broadly U-shaped in transverse profile, and is characterized by a shoal less than 100 m deep at its western end. Submerged U-shaped valleys join the sound at elevations of 200 to 300 m above its floor. These valleys can be traced landward to glacial valleys headed by ice caps (Pelletier, 1964 and 1966). Seismic data collected in 1971 indicate that the acoustic basement lies at relatively shallow depths. Sediments appear to have been deposited within basins formed on either sides of a basement high, the dips of the beds reflecting the underlying structure. Some recent faulting is evident in the sedimentary sequence, and towards the mouth of the sound, major faults with significant vertical displacements are thought to be present (Keen and Barrett, 1973).

At the eastern end of Jones Sound, Glacier Strait and Lady Ann Strait separate Ellesmere, Coburg, and Devon islands. Glacier Strait is shallow, whereas Lady Ann Strait contains an axial trough 500 to 700 m deep which is thought to represent the continuation of Jones Sound (Pelletier, 1964).

2.3 Lancaster Sound

The submarine topography of Lancaster Sound was compiled from a detailed hydrographic survey conducted in 1960 from the Canadian scientific vessel Baffin. Contoured soundings indicate a broad basin increasing in depth from 250 m in the east to 500 m in the west, with a maximum recorded depth of 950 m. Along the northern coast, slopes of 25 per cent were recorded near Dundas Harbour, and these decrease progressively to the west. The southern walls are similar to the northern ones but have more gentle slopes (Buckley, 1971). Pelletier (1966) and Buckley (1971) indicate that

submerged U-shaped tributary valleys join the sound at levels of 200 to 300 m above its floor, and can be traced to glacial valleys on land. Admiralty Inlet and Croker Bay are two such hanging troughs. Both possess a relatively shallow transverse sill, a narrow, slightly deeper channel through the sill, and a U-shaped deep valley behind the sill (Buckley, 1971).

Geophysical and hydrographic studies conducted in 1973 (Lewis et al., 1974) and 1976 (Lewis et al., 1977) indicate that the floor of the sound has an irregular topography. A trough filled with approximately 100 m of sediments traverses the area north of the medial axis of the sound. A second trough filled with less than 90 m of sediments trends northward into the area from Prince Regent Inlet, and runs eastward, south of the medial axis of Lancaster Sound. The bedrock surface rises between the two troughs and forms an elevated axis along which rock outcrops through relatively thin sediments (Lewis et al., 1974).

The predominant sediment type in Lancaster Sound is a mixture of clay and silt with varying amounts of sand and gravel. Gravel, commonly composed of sub-angular pebbles and cobbles, blankets the sea floor adjacent to the coast and medial rise (Lewis et al., 1974). South of Devon Island, sea floor currents appear to winnow the finer material present in the glacial till and leave areas of pavements of pebbles and cobbles.

Magnetic anomaly data collected in 1970 and 1971 support the hypothesis that Lancaster Sound is a graben, as both north and south margins are founded by faults. The presence of a regional northerly dip suggests that subsidence has been greater on the northern side of the sound (Keen et al., 1972a; Keen and Barrett, 1973; and Lewis et al., 1977).

2.4 Baffin Bay

In Baffin Bay, the bottom slopes from depths of 200 to 400 m below sea level in the north where the sea floor is continuous with that of Smith Sound, to depths greater than 2,300 m further south. The entire eastern boundary is formed by the western coast of Greenland, and the western boundaries are discontinuous and consist of the eastern islands of the Canadian Arctic Archipelago. North to south, the bay is approximately 1,300 km long and is 450 km wide in the central area.

2.4.1 Northern Baffin Bay

In northern Baffin Bay, a distinct continental shelf and slope is present, and shows slopes slightly steeper in the east than the west. Submarine topography is hilly and resembles a submerged headland previously exposed to subaerial weathering. This submerged headland drops to a regularly sloping surface approximately 224 km from shore, at depths of about 600 m (Pelletier, 1966; and Keen et al., 1972a). It seems that the continental shelf has been broadened and its slope reduced by the deposition of sediments derived from the Arctic islands and transported through Lancaster Sound and Smith Sound (Pelletier et al., 1974).

2.4.2 The Eastern Margin

The eastern shelf is wide and consists of an inner zone that extends 100 km offshore to depths of 300 m, and an outer zone that is approximately 150 km wide and grades gently to the shelf break. From the shelf break, the continental slope dips steeply to depths of about 2,000 m. The width of the outer shelf varies considerably, and it narrows rapidly south of 71° where the shelf break occurs at 300 m.

The continental shelf is dissected by deep, sinuous valleys that extend from land to the edge of the continental slope. Similarly, drowned headlands or interfluves occur between these valleys and are contiguous with topographic features on land (Pelletier, 1966; and Pelletier et al., 1974). Detailed studies of southwestern Greenland have shown the presence of marginal channels several hundred metres deep, separating the inner and outer zones of the continental shelf. In general, these channels trend parallel to the coastline and may mark a fault or hinge line that probably denotes the transition to Mesozoic and Tertiary sequences. Johnson et al. (1976) suggest that these channels were deepened by ice scouring during glaciation.

Seismic and aeromagnetic surveys across the continental shelf of Melville Bay defined the location and the extent of a graben trending southeast-northwest, parallel to the coastline (Keen et al., 1972b; and Keen and Barrett, 1973). The graben varies between 50 and 60 km in width and appears to be bounded by horsts whose tops are close to the sea floor. It is estimated that about seven km of sediments are present in the faulted basin.

2.4.3 The Western Margin

The submarine geomorphology of the eastern coast of Baffin Island has been extensively reviewed by Loken and Hodgsen (1971). They indicate that the overall width of the continental shelf varies between 35 and 60 km, and becomes wider towards the south. The shelf is dissected by a number of transverse troughs over 600 m deep, that lie offshore from major fiords. These features have also been noted and described by Pelletier et al. (1974); Levy (1979); Maclean et al. (1981); and Maclean (1977, 1981). The deepest troughs are associated with Scott Inlet and Home Bay, where several fiords join. The smooth U-shaped profiles suggest that these troughs were created by glacial scouring. Their size appears to be determined by the width of the ice stream that they channelled. Ridges 30 m high along the sides of Home Bay, Clyde Inlet, Scott Inlet and other troughs have been interpreted as submarine extensions of lateral moraines (Loken and Hodgsen, 1971; and Loken, 1973). Marginal channels present off the glaciated coasts of Greenland, Labrador and Norway do not occur on the continental shelf of Baffin Island. This may be due to either the structural history of the area or its location with respect to the marginal zones of deposition of the Pleistocene ice sheets.

The troughs off Scott Inlet and Buchan Gulf have been the focus of recent geophysical and geological investigations to determine the possibility of natural seepage of petroleum from the seabed (Levy, 1979; Maclean, 1981; and Maclean et al., 1981). Anomalies in petroleum residue have been found in the surface microlayer, water column, and unconsolidated sea floor sediments of both the Scott and Buchan troughs. The most continuous seep was located on the outer part of the Scott Inlet trough, near the south wall where sedimentary strata flank a basement high.

Seismic refraction data obtained from the northern area of the Baffin Shelf show a series of terraces separated by steeper slope elements. The sediments beneath the central terrace appear to be highly disturbed, and the reflection profiles suggest that this may result from large scale slumping related to the relatively high earthquake activity of the area (Keen et al., 1972b). Similar terrace-like features have been noted in the south on the shelf area between Cape Dyer and Broughton Island, and are thought to be indicators of former sea levels. Analysis of the depositional environment, under which shelf sediments of sands and gravels intermixed with mud, also indicates that changes in sea level have occurred. Bottom sediment colour demonstrates reducing conditions, but the colour of subbottom sediments is indicative of oxidizing conditions. It is postulated that during times of maximum glaciation, exposure of the shelf may have produced the oxidation of the sediments. As the sea later transgressed, reducing conditions may have followed (Baker and Friedman, 1973). These sediments are currently affected by iceberg activity. A sidescan survey of the shelf during the summer of 1976 revealed the presence of large areas of scoured bottom in water depths in excess of 250 m (Falconer, 1977).

The steepness of the continental slope varies between two and four degrees with the general slope decreasing from north to south. These slope angles are gentle in comparison to other glaciated coastlines, where they can reach angles of 30°. Although the continental slopes are generally smooth and gentle, small rises with 25° slopes do occur. These may reflect recent crustal movements, or may have been produced by localized slumping (Loken and Hodgsen, 1971).

2.4.4 Central Baffin Bay

The central part of Baffin Bay reaches depths greater than 2,300 m, which is five to ten times the depth of most arctic channels. The sea floor is a featureless abyssal plain which does not appear to have undergone subaerial erosion. The central basin is bounded to the east and west by steep continental slopes, which, in turn, show no evidence of the type of terrestrial erosion present on the adjacent continental shelves (Pelletier, 1966). During the past decade, Baffin Bay has been the focus of intensive geophysical and geological studies in an attempt to define the presence or absence of oceanic crust, and hence, determine its origin. Data obtained from several cruises have identified the presence of oceanic crust in the central part of Baffin Bay as illustrated in Figure 11 (Keen et al., 1972a

and b; Wetmiller, 1974; Pelletier et al., 1974; Hood and Bower, 1975; and Jackson et al., 1977). There is, however, no conclusive evidence for the presence of a buried ridge beneath the bay, nor have fracture zones been delineated. Some conflict surrounds the evidence for the presence or absence of magnetic lineations which are considered indicators of sea floor spreading. Keen et al. (1972a and b) suggest that there is no evidence for or against the presence of such lineations, whereas Hood and Bower (1974) and Beh (1975) confirm their presence in northern Baffin Bay.

2.4.5 Davis Strait

Within Davis Strait, a submarine rise often referred to as the Davis Strait Sill, extends eastward from Cape Dyer on Baffin Island to Greenland. Minimum water depths in the strait are about 400 m. To the north, water depths increase to 2,300 m in Baffin Bay, and to the south, water depths in excess of 3,000 m are present in the Labrador Sea. On the eastern side of the strait, a continental shelf and slope are well developed, whereas they are absent from the western side. There, a wide continental shelf merges gradually with the sill. Topography over the sill is irregular, and is characterized by a surface locally cut by deep channels (Jezeck, 1972; and Wallace, 1973).

Davis Strait is a key area in attempts to understand the origin of Baffin Bay and the Labrador Sea. The landward edges of the Davis Strait Sill are marked by Tertiary basalt volcanic rocks. If they are continuous along the sill, they could represent the site of an upwelling convection plume.

2.4.6 Origin and Evolution of Baffin Bay

There are several theories concerning the origin of Baffin Bay. While there is general agreement that the Eurasian part has formed as a result of spreading away from the Nansen Ridge during the past 40 to 60 million years, there is no consensus concerning the Amerasian part which includes the Baffin Bay region. Geophysical surveys have defined the presence of oceanic crust in Baffin Bay, but have failed to produce conclusive evidence of the presence of magnetic lineations or a ridge.

An early model suggested by Keen et al. (1972b) suggests that during the early Cretaceous, Canada separated from Greenland by opening a part of the Labrador Sea and a small area of Baffin Bay. At this time, the pole of rotation was located on Bathurst Island. During the early Tertiary, the pole of rotation moved to the Pacific Ocean, and plate motion along Baffin Bay was translational along Nares Strait. There are several structural and geological features in Baffin Bay that support this theory. When Baffin Bay and the Labrador Sea are closed according to this model, the Tertiary basalts present on opposite sides of Davis Strait are brought together without overlap. This accords well with the age of the basalts, and is consistent with the hypothesis forwarded by Hood and Bower (1975) that the

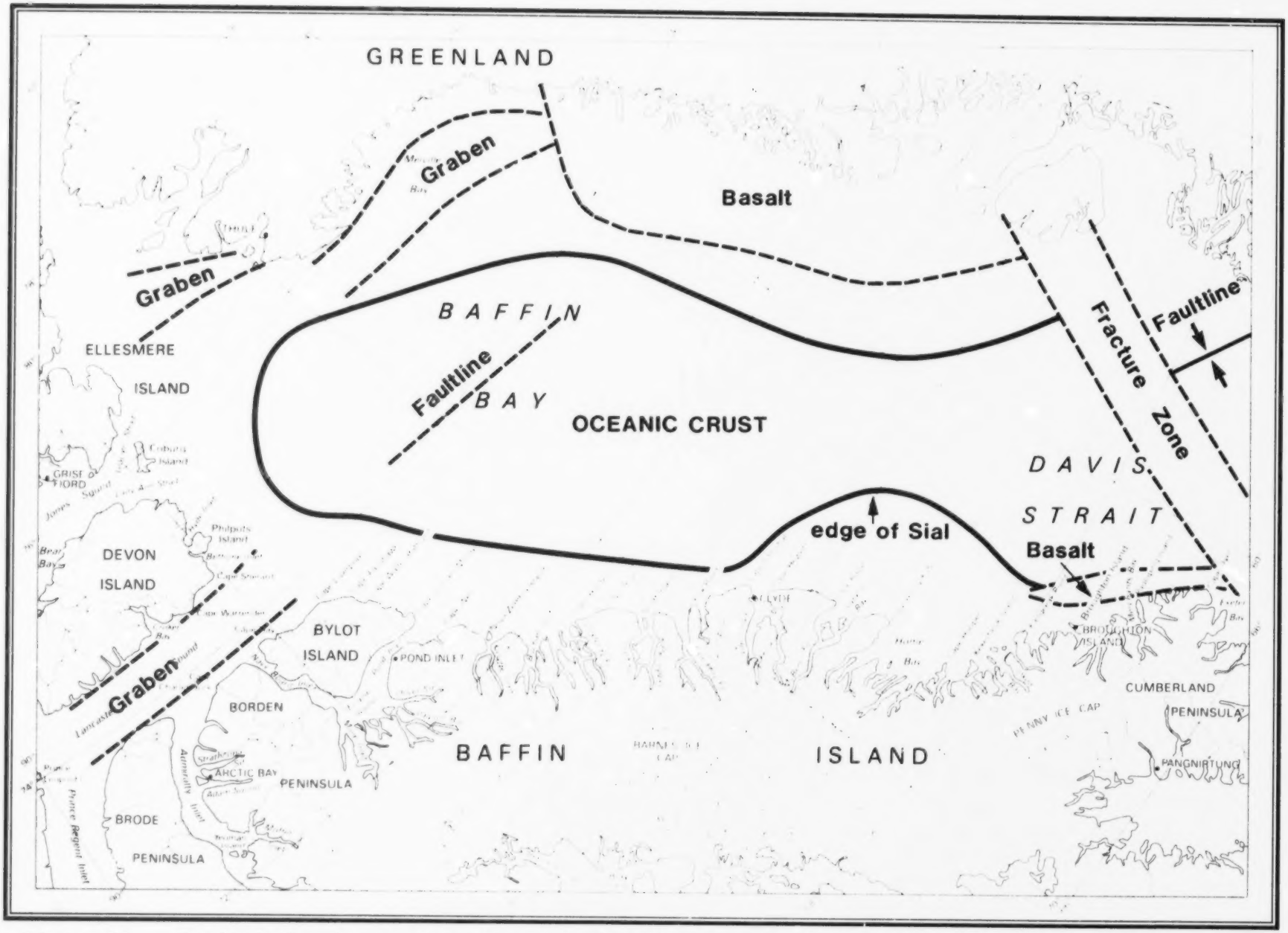


Figure 11 Geological and tectonic regions determined from geophysical surveys, after Hood and Bower 1975.

Davis Strait Sill is a fracture zone. Uplift during the early Tertiary or first phase of opening can account for the mountains of eastern Baffin Island. Relatively undeformed sedimentary rocks underlie much of northern Baffin Bay, and are particularly thick under the northern continental shelf off Lancaster Sound. These sediments could have been derived from the Innuitian region of the archipelago which is thought to have been uplifted during the late Cretaceous and early Tertiary. Eroded material from this landmass could have been channelled to a depositional basin in northern Baffin Bay (Pelletier et al., 1974). Hood and Bower (1975) suggest that as Greenland separated from Canada along transform faults, unstable triple junction faults were formed in the entrance to Hudson Strait, in the vicinity of Disko Island, and in northern Baffin Bay near Lancaster Sound. The Hudson Strait triple junction was formed between a graben in the strait, a transform fault (Figure 11), and the mid Labrador Sea ridge. Movement along this junction may have contributed to the formation of grabens observed in Cumberland Sound and Frobisher Bay. A triple junction in the region of Disko Island could account for a graben-like feature that extends from the island to the mainland. A third triple junction at the northern end of Baffin Bay, between Nares Strait and Lancaster Sound could account for a graben and the presence of a major east-west trending fault in the area. Reconstructions based on the magnetic lineations of the Labrador Sea suggest that the opening of Baffin Bay may be associated with such a junction in the Lancaster Sound area (Jackson et al., 1977). The Melville Bay graben, filled with basalt at its southern end, could represent an aulacogen or 'false start' to the Baffin Bay rafting (Hood and Bower, 1975).

The presence of a sedimentary basin of Proterozoic age across the entrance to Smith Sound would seem to refute the concept of movement along Nares Strait. Similar sedimentary structures in northern Baffin Bay are relatively undisturbed by fractures or faults. This lack of deformation can only be explained by the "sea floor spreading theory", if they were deposited after spreading had ceased, or were moved as a unit of the crust (Keen et al., 1972b; and Keen and Barrett, 1973.) With reference to the geometry of the model, (Keen et al., (1972b) indicate that the Labrador Sea cannot be completely closed without substantial overlap of the continents in northern Baffin Bay. Continuous magnetic anomalies present in Nares Strait suggest that little or no offset has occurred there. Finally, a pre-drift model showing Ellesmere Island and Greenland widely displaced does not agree with the regional geology (Kerr, 1980).

An alternate means of fitting Greenland and Canada is to visualize Baffin Bay as having been formed by the foundering of continental crust (Beh, 1975). Initial uplift would be accompanied by volcanism and rifting, then subsidence of the margin. The margins of Baffin Bay do show evidence of Tertiary volcanism, followed by a period of marine sedimentation during the Cretaceous. Uplift appears to have occurred during the Tertiary as evidenced by the presence of basalt breccias deposited at, or close to, sea level on Disko Island (Jackson et al., 1977; and Keen et al., 1972a and b). However, even if the unusually good fit of southern Greenland into the Labrador coastal notch was ignored, the unusually thin crust in Baffin Bay is the exact opposite to what is expected when continental crust founders.

A compromise theory advanced by Kerr (1980) interprets Nares Strait as a transform structure with minor left lateral strike-slip displacement, minor rotational separation, and major subsidence of continental crust by thinning and foundering.

The origins of the channels of the Arctic Archipelago have been discussed by Fortier and Morley (1956). They suggest that these channels are a submerged and glacially modified fluvial system that drained a continuous land mass during Tertiary time. According to their hypothesis, rivers emerging from Smith, Jones and Lancaster sounds and the east coast of Baffin Island all drained into Baffin Bay (Pelletier, 1964, and 1966). During the Pleistocene, glaciers followed the pre-existing drainage system and modified the landscape to create fiords along the eastern coast of Baffin Island, and a deeper and smoother profile in Lancaster and Jones sounds. While glacial features are present in the major sounds and along the Baffin Bay margins, it is difficult to reconcile the concept of a single Tertiary landmass having a drainage system with the present pattern, and the concepts of sea floor spreading in approximately the same time period.

3 Tectonic Geology

Tectonics is a branch of geology that deals with the broad architecture of the outer part of the earth; that is, the regional assemblage of structural or deformational features, their mutual relationship, origin, and history of evolution. A tectonic map portrays the architecture of the outer part of the earth by showing not only structures, but also ages and types of rocks on which structures are superimposed (Bates and Jackson, 1980).

Most of the eastern Arctic is made up of a part of the Canadian Shield, a vast region of Precambrian rocks that occupies nearly half of Canada. The Shield itself has been subdivided into seven structural provinces and 34 subprovinces according to differences in structural trends and styles of deformation (Stockwell, 1982; and Douglas 1980). Different orogenies or periods of deformation, including thrusting, folding, and faulting in the outer and higher layers, as well as plastic folding, metamorphism and plutonism in the inner and deeper layers, account for major differences in structural trends. Provinces and subprovinces differ from orogenic belts in that they contain not only orogenically deformed rocks, but undisturbed cover material and a variety of post-orogenic intrusions.

Relative ages between orogens can be observed in the field by angular unconformities and by the truncation of older structures by younger ones. With the advent of isotopic or radiometric dating, that is, the measurement of absolute geological time by the study of parent and daughter isotopic abundance, it is now possible to confirm observed relative ages, and to outline the geographical extent of different orogenies. Radiometric dates are unfortunately limited by large, inherent analytical errors, and orogenies cannot yet be defined accurately in absolute time. To circumvent this problem, orogenies are defined by concrete rocks in type regions rather than by absolute dates. Type regions serve as standards for reference, and any improvement in their dating or mistake in correlation do not invalidate the units.

Following this scheme, four orogenies have been defined in the Shield. The Kenoran Orogeny is defined as the last important period of widespread folding, metamorphism and plutonism in the Superior Province, which is chosen as the type region. Similarly, the Hudsonian Orogeny has been defined in Churchill Province, the Elsonian Orogeny in the Western Main Province and the Grenvillian Orogeny in the Grenville Province. In each province, radiometric dates show a considerable spread, but displayed as a histogram, they fall in distinct groups that indicate the dominant orogeny. The time interval from beginning to end of any orogeny is difficult to determine, mostly because of analytical errors, but as these tend to cancel out, each date group yields a significant mean value. The mean date value for the four orogenies in their type provinces are: the Kenoran, 2,500 m.y.a.; the Hudsonian, 1,750 million years ago (m.y.a.); the Elsonian, 1,400 m.y.a.; and the Grenvillian, 1,000 m.y.a.

As orogenies are time units, the names used in the type regions are used to describe contemporaneous orogenies even if widely separated geographically. Several of the provinces have been involved in orogenies older or younger than the dominating one. This is why a certain amount of correlation exists between orogenies in different provinces. Part of the Churchill Province makes up most of the eastern Arctic. For this province, a histogram of dates shows a major grouping corresponding to the Hudsonian Orogeny, a minor one corresponding to the Kenoran Orogeny and essentially none for the Elsonian and Grenvillian Orogenies. Sometimes, the effects of the dominating orogeny are so strong that no older dates survive. In each province, it is the last dominating orogeny that is responsible for its characteristic structures.

The four dated orogenies, each followed by an erosional interval provide the basis for the time-stratigraphic division of the Precambrian as follows:

Eon	Era	Sub-Era	Orogeny Mean K-Ar mica age in million years
			(750)
	Hadrynian		
			Grenvillian (1,000)
		Neohelikian	
			Elsonian (1,400)
Proterozoic	Helikian	Paleohelikian	
			Hudsonian (1,750)
	Aphebian		
			Kenoran (2,500)
Archean			

The first units are called Eons. The Proterozoic and Archean Eons allow a gross time classification of poorly-defined rock units. The Archean Eon includes all rocks involved or emplaced during the Kenoran Orogeny as well as all older rocks (Stockwell, 1976). The Proterozoic is subdivided into three Eras: the Aphebian, Helikian and Hadrynian, with the Helikian further subdivided into two sub-Eras, the Paleohelikian and Neohelikian, each of these terminating at the end of the corresponding orogeny, while the Hadrynian terminates at the start of the Cambrian.

The gneissic terrain of most of Baffin Island consists of Archean metamorphic rocks, mostly granitoid in composition and lower to upper amphibolite in facies. These are thought to have been emplaced or metamorphosed about 2,800 m.y.a. Kenoran orogenesis, granodioritic plutonism and metamorphism affected most of the Archean rocks approximately 2,555 m.y.a. By early Aphebian time, most of the Archean terrain had been metamorphosed to the amphibolite facies, under moderate to high pressures, and temperatures in the range of 450 to 700° C. In this metamorphic facies, basic rocks are characterized by hornblende and plagioclase, and pelitic rocks by mica, almandine, staurolotite, and kyanite or sillimanite. Following a period of uplift and erosion, three major Lowe Aphebian sequences were deposited in a shelf, miogeoclinal (a prograding wedge of shallow water sediments at the continental margin), and eugeoclinal (continental rise) environments in central, southeastern and southwestern Baffin Island. Granitic plutons were emplaced from 2,000 to 1,900 m.y.a. and the area was subjected to the intense Hudsonian orogenesis and metamorphism between 1,964 and 1,600 m.y.a. After the Hudsonian deformation ceased, metamorphism continued to the granulite facies in several scattered locations on Baffin and Bylot islands. The facies is typical of high pressure, deep-seated regional metamorphism at temperatures in excess of 650°C. In this facies, basic rocks are characterized by diopside, hypersthene, and plagioclase; and pelitic rocks by sillimanite or kyanite, perthitic feldspar, almandine, and often cordierite. Many massive quartz monzonitic and monzocharnockitic plutons are present in central southern Baffin Island. Some of these may have melted and recrystallized in situ, and most charnockites are thought to include a considerable amount of Archean and Aphebian granitic material. As much as 20 km of uplift may have occurred after the Kenoran Orogeny, and up to 30 km of uplift in parts of Baffin Island following the Hudsonian Orogeny. The Baffin Island crust was probably 60 to 70 km thick at some time before the Helikian, and it is thought to be 32 km thick at present (Jackson et al., 1977). Neohelikian strata, and possibly Hadrynian diabase dikes, have been modified by low grade metamorphism with temperatures of up to 300°C. Phanerozoic strata have not undergone this subgreenschist metamorphism. Faulting has been active since the Helikian, and has divided Baffin Island into three major blocks tilted upward towards the northeast (Jackson and Morgan, 1978).

4 Geology of the Eastern Arctic

Geology is the study of our planet, the material of which it is composed, the processes that act on these materials, the products formed, and the history of the planet and its life forms since its origin. Geological studies in the eastern Arctic began with early explorers and continue today. The Geological Survey of Canada has conducted numerous field studies, and the results have been published in many journals.

4.1 Geology of Southeastern Ellesmere Island

The basement rocks of southeastern Ellesmere Island form two distinct groups: dark feldspar-quartz gneisses of granitic aspect to the west of King Edward VII Point; and quartzites, gneisses, and crystalline limestones of metasedimentary aspect to the east of the same Point. Both groups include leucogranite or pegmatite-aplite bodies in variable proportions.

The dark gneisses contain quartz, plagioclase, and minor amounts of biotite. The weak foliation is due to the parallel orientation of patches of biotite crystals, and the gneissosity is due to the segregation of large feldspar crystals. Most of these gneisses are medium grained and some have an equigranular texture. White quartz feldspar veins, dikes, and irregular bodies are present in small proportions and augen structures are present locally.

Coburg Island and the area east of King Edward VII Point are underlain by a greater variety of gneissic rocks. Some of these gneisses are similar to those described above, but garnetiferous rocks and crystalline limestones are also present. Locally up to 50 per cent of the rock is composed of white pegmatite that occurs in irregular sills and dikes. These pegmatites contain quartz, plagioclase, some potassic feldspar and not uncommonly, garnet, biotite and muscovite.

Between Wilcox and Cory glaciers, leucogranite forms a large proportion of the rock. As opposed to the white pegmatites, these leucogranite sills often contain minerals that are different from those of the host rocks. These bodies may mark a zone of intrusion or metasomatism coinciding with the westward margin of the metasedimentary terrain east of King Edward VII Point.

The two Precambrian lithological regions have contrasting structural trends. In the eastern metasedimentary terrain, bedding and foliation predominate and trend north to northwest. In the western granitic terrain, foliation is variable but trends mainly eastward.

A discontinuous outcrop of crystalline limestone appears to be a remnant of a bed or a zone of beds, and forms the striking Princess Monument at the southeastern tip of Coburg Island. A few diabase dikes five to thirty m wide, are widely distributed and trend in general, north or northeast.

Early Paleozoic limestone, dolomite and sandstone unconformably overlie the basement gneisses. Inliers of gneisses and scattered outcrops of Paleozoic rocks result from the gentle northward to westward dip of the unconformity, the presence of faults, and the highly dissected nature of the terrain. Faults trending easterly to northerly cut all bedrock units. Paleozoic strata along the southern coast dip gently northerly, perhaps as a result of the tilting of fault blocks. On the extremity of South Cape, strata form closely spaced, east-trending folds. Dips are variable and generally moderate. These folds are conspicuous in a region of broad and gently inclined structures (Christie, 1961; and Frisch and Morgan, 1978).

4.2 Eastern Devon Island

On the northern coast of Devon Island, Precambrian rocks consist mainly of quartz-feldspar-biotite gneisses with weak to strong foliation, generally trending east. These gneisses can be divided into two groups: weakly foliated granitoid and strongly foliated garnetiferous gneisses. Dark grey, weakly foliated, granulose gneisses and cataclastic textures occur frequently in both groups.

The granitoid gneisses occur between Sverdrup Inlet and Cape Sparbo, and are usually medium to coarse grained. Occasional pegmatitic veins have essentially the same mineralogy as their host rocks, and in places have gradational contacts. East of Cape Hardy, the weakly foliated gneisses give way to a grey, well - foliated to - banded gneiss and a granitoid rock. Mylonitic and cataclastic textures are more abundant to the east. In this region, the foliation parallels the regional gneissic trend.

The second major group, the garnetiferous gneiss, is exposed in the valley floors south of Cape Newman Smith. There, the gneiss appears to be of metasedimentary origin, and trends easterly to southeasterly. Cataclastic textures are common south and east of Belcher Point.

On the eastern coast of Devon Island, the Precambrian rocks consist of granitoid quartz-feldspar gneisses, subgranitoid foliated gneisses and foliated or banded garnetiferous quartz-feldspar gneisses. The variety is due to the coastline cutting across the prevailing easterly structural trend. A zone of strongly foliated garnetiferous gneiss outcrops south of Cape Fitz Roy, and underlies the northern shore of the deep inlet. The coast between the inlet south of Cape Fitz Roy and Cape Parker consists of granitoid gneisses with a weak foliation. South of Cape Parker, the gneisses are well-banded, mainly leucocratic and garnetiferous. At Hodgson Head, they become quartzitic, with flat-lying banding and widespread pegmatitic intrusions. On Philpots Island, they vary between strongly foliated to banded in the north, and grey, near massive in the south. The coast between Philpots Island and Cape Sherard consists mainly of outcrops of light to dark grey or brown biotitic, granulose, quartz feldspar gneisses.

On the southern coast of the island, the Precambrian rocks consist of foliated, banded or uniform, cataclastic gneisses. These are well-exposed between Cape Sherard and Croker Bay. East of Dundas Harbour, the gneiss is often dark, with boudins, inclusions, breccia zones and ptygmatic folds. To the west of Dundas Harbour, they become leucocratic and often show lenses of biotite or garnet. Along the southern coast, the gneissic banding and foliation trend easterly or northeasterly with considerable local variations.

Two types of dikes occur on Devon Island and are abundant in some places, especially on the northern side of Philpots Island and south of Cape Sparbo. They are pre- or sin-kinematic, dioritic dikes and post-kinematic diabase dikes. Both trend generally east to southeast. They often have sharp contacts, sometimes with minor extensions into the host rock, and they often show a porphyroblastic texture.

Faults are conspicuous in the Paleozoic beds on eastern Devon Island, and lineaments, presumed to be faults, are present in the Precambrian terrain. Most of these features trend east, and many are sinuous. In general, eastern Devon Island can be described as a horst-like basement partially capped by a nearly undisturbed, thin veneer of Paleozoic strata (Christie, 1978).

Paleozoic strata occur on eastern Devon Island in a weakly west-dripping homoclinal succession, such that Cambrian and Ordovician formations occur to the east and Ordovician, Silurian and possibly Devonian formations occur to the west. At Dundas Harbour, the Precambrian rocks are unconformably overlain by the Lower Cambrian Rabbit Point Formation. Above this are the Middle Cambrian Bear Point and Ooyahgah formations, the Lower Ordovician Mingo River and Nadlo Point formations, and the Middle Ordovician Croker Bay Formation. At both Burnett and Sverdrup inlets, the Precambrian basement is overlain by at least 400 m of strata, the top of which is probably Ordovician (Fortier et al., 1963).

4.3 Brodeur and Borden Peninsulas

The oldest units on Brodeur and Borden peninsulas consist of Aphebian granitic and gneissic rocks. On northern Borden Peninsula, crystalline rocks extend east from the ice cap at the head of Elwin Inlet to Navy Board Inlet and include greyish or pinkish medium to coarse grained plagioclase-quartz-microcline gneisses with variable proportions of ferromagnesian minerals. On southern Borden Peninsula, radiometric dating indicates an Aphebian age. Yeoman Island consists of a dated Aphebian crystalline complex (Blackadar et al., 1967b).

The oldest Helikian unit, the Nauyat Formation, consists of relatively unaltered basic flows, massive basaltic rocks and amygdaloidal basalts. On northern Borden Peninsula, basalt flows are dark in colour and have an estimated thickness of about 300 m. More than 350 m of this type of rock is exposed on southern Borden Peninsula, south of Adams Sound (Jackson et al., 1978).

The Adams Sound Formation, the youngest member of the Eqalulik Group, consists of pale orange silica-cemented sandstone. The total thickness of this unit appears to exceed 300 m, especially in southern Borden Peninsula where it may exceed 1,200 m. A gradational contact separates this unit from the overlying Arctic Bay Formation.

The type section of the Arctic Bay Formation, the oldest member of the Uluksan Group, consists of 120 m of black, fissile, medium to fine grained argillaceous limestone and calcareous dolomite. On structural grounds, a thickness of about 600 m has been postulated along Alfa River and east of Milne Inlet (Blackadar et al., 1967a).

The type section for the Society Cliffs Formation measured near Arctic Bay includes about 300 m of vaguely bedded, light-grey dolomite. This formation becomes more argillaceous eastward, and is separated from overlying units by conformable contacts.

A succession of dark to pale grey, medium to coarse grained dolomite, conglomerate and interbedded limestone and mudstone forms the next unit, the Victor Bay Formation. East of Tremblay Sound, this formation includes 450 m of strata. Near Arctic Bay, the type section is only 150 m thick. Near Victor Bay, this formation is conformably overlain by the Athole Point, elsewhere, the Strathcona Formation forms the overlying strata.

A gradational contact separates the above and the Elwin Formation. This formation consists of an assemblage of alternating reddish or orange quartzitic or silty sandstone and micaceous shales. The thickness of the formation exceeds 150 m.

Basic dikes similar to those found on northern Baffin Island intrude all strata older than Paleozoic ones. Some of the northwest trending dikes exceed 150 m in thickness, and most of the north trending ones are narrower.

Paleozoic strata rest with a slight angular unconformity on several Helikian formations, and with a pronounced unconformity on Aphebian rocks. The Paleozoic record includes the Admiralty Group, mainly Cambrian and Early Ordovician fluviatile or intertidal formations, of which some may be correlative with those of eastern Devon Island. This group has a thickness that is variable from one location to another, but that may exceed 600 m.

The next unit consists of the Middle and possibly Lower Ordovician Ship Point Formation. It is separated from underlying strata by a paraconformity that may represent a major hiatus during the Upper Cambrian. The formation consists of a fine grained to microcrystalline dolomite, and has a thickness of 100 to 300 m.

The Brodeur Group overlies the above, and consists of dark cryptocrystalline limestone, dolomite and brecciated evaporites and stromatolitic beds. This group appears to be present only on northern Brodeur Peninsula.

On Northern Brodeur Peninsula, Paleozoic rocks form part of an arcuate structure, the Brodeur Homocline, and in general, dip gently towards the northwest. In the southwestern part of the peninsula, these strata are generally horizontal. On the northern coast of Borden Peninsula, a belt of strata dips gently north, and to the northwest, they are essentially horizontal. In central Borden Peninsula, these rocks are also horizontal, but they are locally disturbed by groups of faults. In the southern part of the peninsula, easterly trending faults mark the transition between Paleozoic and Precambrian rocks (Blackadar et al., 1967c).

An important lead-zinc mine, The Nanisivik mine, is located on the southern side of Strathcona Sound near Arctic Bay. The mine was brought into operation in 1976 and currently produces 150,000 metric tonnes of ore per year. It has a life expectancy of about 15 years. At this mine, ore is extracted throughout the year, concentrated and stockpiled for summer shipment. The development of this mine has required the construction of a large air strip, a town for several hundred people, and a marine terminal on Strathcona Sound.

4.4 Bylot Island

Highly deformed and metamorphosed rocks underlie most of northern, eastern, and central Bylot Island. Most of these rocks are Aphebian in age and some are Archean. It appears that the Hudsonian orogeny was the last major deformational and metamorphic event in the area. Approximately two-thirds of this complex is composed of banded gneisses that may be metasedimentary in origin. The bands vary from centimetres to more than 30 metres in thickness, and have contrasting compositions. Some are quartzofeldspathic and of metasedimentary origin, and some granitoid, granitic, and pegmatitic rocks appear to be the product of in situ recrystallization and metasomatism. Some bands are rich in biotite, hornblende, and pyroxene. Some areas consist of migmatites that contain a large amount of pegmatite and aplite. Monzocharnockite forms a massive pluton of over 2,500 km^2. A few small outcrops of this so-called Bylot Batholith occur near Cape Hay. This rock type is generally intensely deformed and characterized by lenticular quartz, stretched feldspar augens, and sometimes by a porphyritic texture. Minor anorthositic and mangeritic rocks appear to be genetically related to the monzocharnockite. A few areas in northern and western Bylot Island consist of foliated quartz monzonite that resembles the monzocharnockite.

Extensive uplift and erosion followed the Hudsonian Orogeny and preceded the deposition of the Eqalulik and Uluksan groups in Helikian time. These groups appear to have a thickness of over 2,600 m and they appear to have been deposited in a platform environment.

Most of the overlying Adams Sound Formation consists of a red bed sequence, with the lower member a fluvial deposit, and the upper member a shallow platform deposit. In some locations, this formation consists of a clean sandstone.

The Arctic Bay Formation overlies the Adams Sound Formation conformably. This shallow platform deposit grades upward from sandstone, siltstone and shale to clastics and carbonates. Such a gradation could represent the gradual erosion of the source of the clastic sediments.

The Society Cliffs Formation forms the conformable overlying unit. It consists mainly of dolomite deposited in a shallow carbonate platform environment, and lenses of shales and gypsum produced in evaporitic tidal flat and lagoonal environments.

The Victor Bay Formation is characterized by conglomerates and breccias.

Helikian rocks are gently folded with northwesterly trending axes. Block faulting is present, and may have caused solution breccias in the Arctic Bay, Society Cliffs and Victor Bay formations. Late Precambrian diabase dikes intrude all older rocks. These are part of the Franklin dikes, a vast swarm of dikes that extends in a giant arc from Great Bear Lake to Baffin Island. They are 700 million years old and must have been intruded after the deformation of the Helikian strata (Fahrig, 1971).

Phanerozoic strata, i.e. rocks including and younger than the Cambrian, are well exposed along the shores of Eclipse Sound. The Eclipse group has an estimated thickness of over 1,000 m, and strata of this group are gently folded along northwesterly trending axes.

No mineral concentrations of economic interest have been found on Bylot Island. There are a few locations where magnetite, malachite, siderite, gypsum and coal are present (Jackson et al. 1975).

4.5 Northern Baffin Island

Most of northern Baffin Island is underlain by Archean and Aphebian granitic gneisses and migmatites cut by granitic or basic dikes and sills. A few small gabbro and quartz monzonite plutons were intruded before the Hudsonian Orogeny and some larger bodies grade into migmatite, but have recognizable metasedimentary and metavolcanic origins. Small plutons and dikes of granite or granodiorite are unmetamorphosed and are probably late tectonic. Although all these rocks yield Hudsonian ages, many bodies show a similarity with Archean counterparts in the Icebound Lakes area (Jackson et al., 1975).

The Helikian record consists of 3,000 m of gently folded, unmetamorphosed sediments of the Eqalulik and Uluksan groups. In northern Baffin Island, the Eqalulik Group is represented by the Adams Sound Formation, and the Uluksan Group by the Arctic Bay, Society Cliffs, Victor Bay, and Athole Point formations. These rest unconformably or are in fault contact with older rocks.

The Hadrinian record consists of northwest trending diabase dikes, part of the Franklin dikes, and the Phanerozoic record consists of the Eclipse Group, or about 1,200 m of strata.

Northern Baffin Island may comprise deposits of economic interest. The Society Cliffs, as well as the underlying Arctic Bay and overlying Victor Bay formations, may contain deposits of sphalerite, galena, silver, chalcocite, chalcopyrite, pyrite, marcasite and fluorite. Gypsum and coal beds outcrop at several localities. Recently, an estimated 130 million metric tonnes of iron ore has been located south of Milne Inlet. The development of this Mary River deposit would have considerable consequences, as it would require the establishment of a community, a marine terminal, and a means of transporting the ore.

4.6 Central Baffin Island

In central Baffin Island, rocks have been divided into two major groups: 1) the Mary River Group which extends discontinuously from Scott Inlet, west to the Barnes Ice Cap and Grant-Suttie Bay on the western coast of Baffin Island, and north to Eclipse Sound; and 2) the Piling Group which extends south of Scott Inlet and the Barnes Ice Cap, to northeast of Nettilling Lake and the head of Cumberland Sound.

Radiometric dating of the gneiss basement that underlies the Mary River Group has yielded ages of 2,552 $\pm$ 42 million years, but it is likely that these rocks existed 2,800 million years ago (Wanless and Loveridge, 1978; and Jackson, 1978). Rocks from the Mary River Group have yielded ages between 2,100 and 2,600 million years with a grouping in the 2,100-2,200 range. In terms of regional extent, it is possible that the rocks in the Mary River area (south of Milne Inlet) may be younger than those of the Grant-Suttie Bay area on the western coast of Baffin Island.

In terms of petrology, rocks of the Mary River Group include intermediate to basic metavolcanics, basic and ultrabasic intrusions, a variety of metasedimentary rocks, and a few acid metavolcanics (Jackson and Taylor, 1972). The metasedimentary rocks occur as lenses of biotite-muscovite-quartz-feldspar paraschists and paragneisses in the basal unit of the group. These basal layers also include iron formations. These iron beds are generally fine-grained, and composed of thin, alternating layers of magnetite, quartz and hematite. Nearly pure magnetite and hematite layers as wide as one m are not uncommon. The iron formations are overlain comformably by the upper formations of the Mary River Group, metamorphosed shales and greywackes.

In 1961, Andrews and Ives (Andrews et al., 1971) discovered a very restricted outcrop of sedimentary rock lying unconformably on the Precambrian gneiss about 40 km northwest of the Barnes Ice Cap. The sedimentary rock consists of a thinly laminated, impure limestone. Palynomorph studies revealed the presence of 74 identifiable angiosperm and

pteriophyte specimens that indicate an age no older than Cretaceous or possibly Tertiary. The nearest deposit of approximately similar age lies near Cape Dyer in the southeast, and in the Queen Elizabeth Islands in the northwest. Its presence in a hilltop position suggests a post-depositional faulting and a well-developed drainage system.

South of the Mary River Group, the rocks of the Piling Group outcrop from east of Steensby Inlet and south of the Barnes Ice Cap to the northern part of the Penny Ice Cap (Jackson and Morgan, 1978; and Jackson and Taylor, 1972). The Piling Group rests unconformably on a gneissic basement and the only dates available for this group are 1,665 to 1,805 million years. It is thought that rocks of this group were deposited during early Aphebian time and have undergone orogenesis during the Aphebian. They are characterized by metasedimentary rocks in which the bedding is usually well preserved.

The Piling Group consists of quartzites, marbles, quartz-rich paragneisses, amphibolites and greywackes. Massive quartz monzonitic and charnockitic bodies occur as dikes, sills, irregular masses, and as a massive group that occupies tens of thousands of square kilometres in southern Baffin Island. Most of these show evidence of late Aphebian high grade metamorphism (Wanless, 1970) and some may have recrystallized in situ. The area of Home Bay has been studied in detail by Tippett et al., 1980). In this area, the basement complex forms the cores of most major structures, and probably formed a substratum during the deposition of the Piling Group.

4.7 Southern Baffin Island

The Hoare Bay Group, a thick succession of intensely-deformed metasedimentary and intermediate to basic volcanic rocks of probable Aphebian age, outcrops in most of the southern part of Cumberland Peninsula (Jackson, 1971; and Jackson and Taylor, 1972). Lithologically, these rocks resemble those of the Mary River Group more closely than the Piling Group. They are thought to have been metamorphosed to the amphibolite facies, and include mica-hornblende-feldspar-quartz schists and gneisses, amphibolites, and biotite-hornblende schists and gneisses (Jackson and Morgan, 1978).

Dating of these rocks has yielded ages of 1,721 to 1,740 million years and Pidgeon and Howie (1975) have reported an age of 1,900+-20 million years for some charnockites. It is probable that the Hoare Bay Group has undergone late Aphebian orogenesis.

Small patches of Tertiary basalts occur intermittently along the coast, near and northwest of Cape Dyer. These rest directly on the Precambrian basement or on a thin sequence of sediments (Clarke and Upton, 1971). Where present, this sediment consists of unconsolidated sand, impure sandstone, shale and conglomerates that include features suggesting a terrestrial origin. Their fossil flora indicates a Paleocene age. These restricted deposits are important because they show similarities with the basalts of western Greenland. This suggests that western Greenland and Baffin Island may be fragments of a formerly continuous basalt plateau now separated by Baffin Bay.

5 Glacial Geology

The area of eastern Baffin Island, eastern Devon Island, and southeastern Ellesmere Island is characterized by glacial processes that have modified the terrestrial environment throughout the Pleistocene and Holocene, and that continue to be active today. The present landscape of eastern Baffin Island and the southeastern Queen Elizabeth Islands is dominated by the effects of glacial processes, either directly or indirectly. The processes are active at the present time, either adjacent to ice margins or in terms of fluvio-glacial processes and glacial isostatic adjustments. Ice-free upland areas are characterized by nunataks, cirques, and U-shaped valleys. Moraines, deltas, sandurs, and raised beach sequences, as well as drift or till deposits, abound throughout the area. Although direct glacial processes are important in the formation of the contemporary landscape, this region is also affected by glacio-fluvial, periglacial, coastal and marine processes.

5.1 Chronology

The Quaternary Period was characterized by large climatic changes, successive and extensive glacial events, and by the development of Man. This period, covering the last two to three million years of the earth's history, includes both the Pleistocene Epoch and the short, non-glacial Recent or Holocene Epoch.

During the Pleistocene, approximately 30 per cent of the surface of the earth and 97 per cent of Canada was glaciated. At present, less than 10 per cent of the earth and 1 per cent of Canada remain under glacial ice (Prest, 1976).

The main events of the Quaternary and their sequence is shown in Table 2. The traditional concept maintains that four main glaciations, the Nebraskan, Kansan, Illinoian, and Wisconsin, occurred during the Pleistocene Epoch, each event lasting approximately 100,000 years. These glacial events were separated by the longer Aftonian, Yarmouthian, and Sangamonian interglacial stages. During these stages, the climate became similar to, or warmer than the present regime, and with associated glacial retreats, the continental areas became largely free of ice. The major glacial advances also included short, non-glacial episodes named interstades. During the interstades, with increased temperatures, glaciers receded extensively but did not disappear from the land masses.

One major interstade recognized in southern Canada is used to subdivide the last glaciation into Early and Late or Classical Wisconsin stages. The term Foxe Glaciation is used to describe the Classical Wisconsin's maximum in the eastern Arctic.

There exists a considerable variation in levels of knowledge throughout the region. Certain locations have been investigated in considerable detail, particularly northern Cumberland Peninsula and Clyde on Baffin

Period	Epoch	Stage		Stade		Approximate date in years × 10^3
		Glacial	Interglacial	Glacial	Interglacial	
Quaternary	Recent or Holocene					0–7
	Pleistocene	Wisconsin		Late or Classical Wisconsin		22
					Mid-Wisconsin	55
				Early Wisconsin		110
			Sangamonian			235
		Illinoian				375
			Yarmouthian			
		Kansan				
			Aftonian			
		Nebraskan				
Tertiary						2,000 to 3,000

Table 2 Chronology of major glacial events of the Quaternary Period.

Island. Other regions have only been studied at a reconnaissance level from aerial photography. This discrepancy in the level of detail of the data causes problems for interpretation on a regional level, as it is frequently not possible to relate events and glacial features from one site to adjacent regions (Nelson, 1980).

Many complex changes have taken place during and since the Pleistocene, on very short time scales and over small geographic areas. Major glacial advances and retreats have, in some cases, taken place over time spans in the order of decades or centuries only. In some areas, the present limits of glaciers are within a few kilometres of their maximum glacial advance. The complexity of the glacial history, combined with the short time span of some of the changes and with the small geographical areas that are involved, lead to a regional understanding that is largely speculative. Although the general trends over the past 100,000 years can be reasonably identified and described, there is no understanding of the details of events and the resulting glacial landforms since the onset of the last glaciation.

A general time scale since 120,000 years BP (before present) for two areas of eastern Baffin Island is given in Table 3. The timing of Clyde River events is shown in more detail in Figure 12 (Column 4) and that for northern Cumberland Peninsula in Figure 13 (Column H). A more detailed chronology for the past 10,000 years is presented by Miller (1973) in Figure 14.

5.2 History of Exploration

The glacial history of the eastern Arctic, including the eastern coastline of Ellesmere, Devon, Bylot, and Baffin islands was largely unknown until recent (1962) geographical and geological surveys were conducted in the area. Early research expeditions (the Wordie Expedition in 1934, and the Arctic Institute of North America Baffin Island Expedition in 1950) delineated the presence of an extensive ice cap on Baffin Island (the Barnes Ice Cap) and indicated that the Baffin Island area could provide an important laboratory for studying active glaciation. The acquisition of air photos for the entire Baffin Island area during the 1940's by the Royal Canadian Air Force provided an additional data source for examining the glacial history of Baffin Island; however, major research programs were not initiated until the early 1960's. The first large-scale field program, the Barnes Ice Cap Survey of 1962, was initiated by the Geographical Branch of Canada. Studies focussed on conducting reconnaissance mapping, as well as identifying active processes and recently deposited glacial sediments in central Baffin Island. Additional detailed glacial studies have focussed on the recent Wisconsin glacial history of eastern Baffin Island (sponsored largely through the Institute of Arctic and Alpine Research), and the glacial history of northern Baffin Island, Bylot Island and southern Ellesmere Island (sponsored primarily through the Terrain Sciences Division of the Geological Survey of Canada). As a result of the recent research, there is a large volume of information on the glacial history of Baffin Island;

CLYDE AREA		CUMBERLAND PENINSULA	
(1) HOLOCENE			
retreat	present	retreat	
	75 BP		
	360		
neoglacial advances	790	neoglacial advances	
	1,660		
	3,200		
hypsithermal warm interval	4,000 - 8,000	hypsithermal warm interval	
Sam Ford Stade	8,000 - 10,000	Cockburn Substage	
(2) WISCONSIN (Laurentide/Foxe Glaciation)			
11,000 - 65,000	Quajon interstade	11,000 - 45,000	
65,000 - 80,000	Ayr Lake stade	45,000 - 55,000	Napiat stade
80,000 - 90,000	Cape Broughton interglacial	55,000 - 75,000	Quajon interstade
		75,000 - 115,000	Alikdjuak stade
90,000 - 115,000	Clyde stade		
		115,000 - 120,000	Cape Christian interglacial
115,000 - 120,000	Cape Christian interglacial		

Table 3 Glacial events of eastern Baffin Island, after Andrews and Barry, 1978; and Andrews et al. 1974.

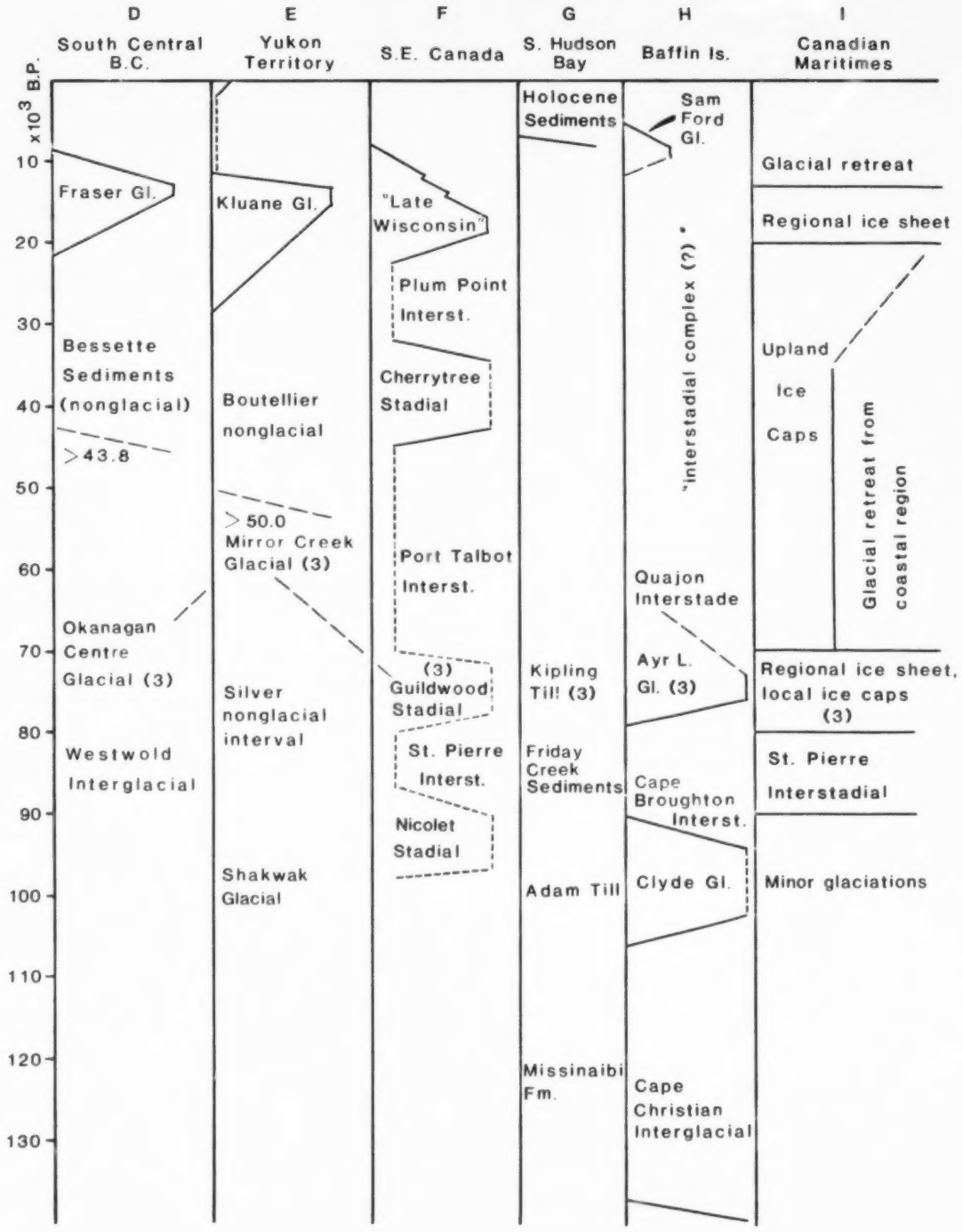


Figure 12 Wisconsin glacial chronology of the Cumberland Peninsula area of Baffin Island, after Andrew et al. 1974.
1-4: eastern Baffin Island, 5: central east Greenland, and 6: paleoprecipitation curves for Camp Century ice core, Greenland.

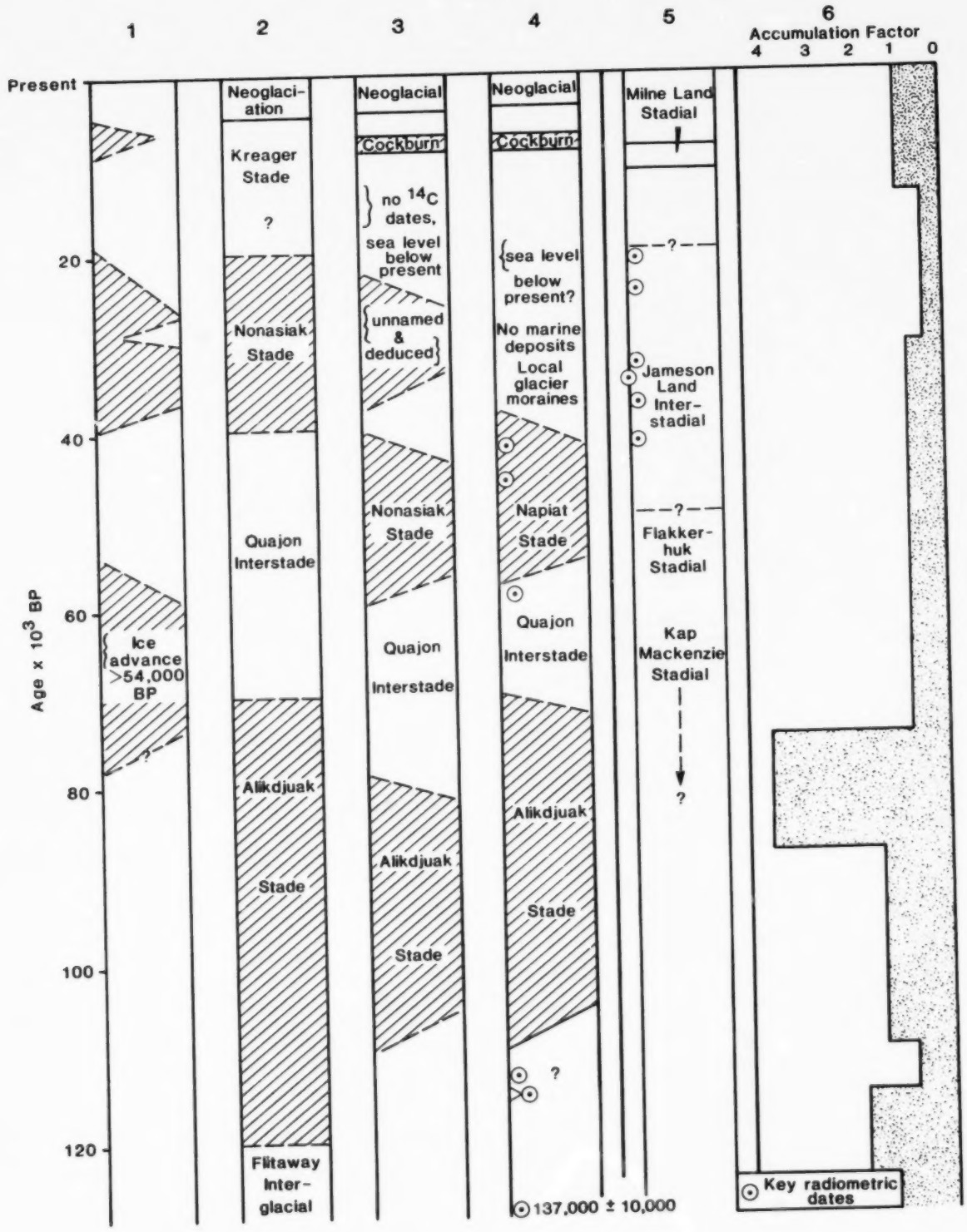


Figure 13 Regional stratigraphic sequence from the Clyde area of east central Baffin Island, after Andrews and et al. 1974 Scale values of X axis of column 6 are factors of current snow accumulation increasing to the left.

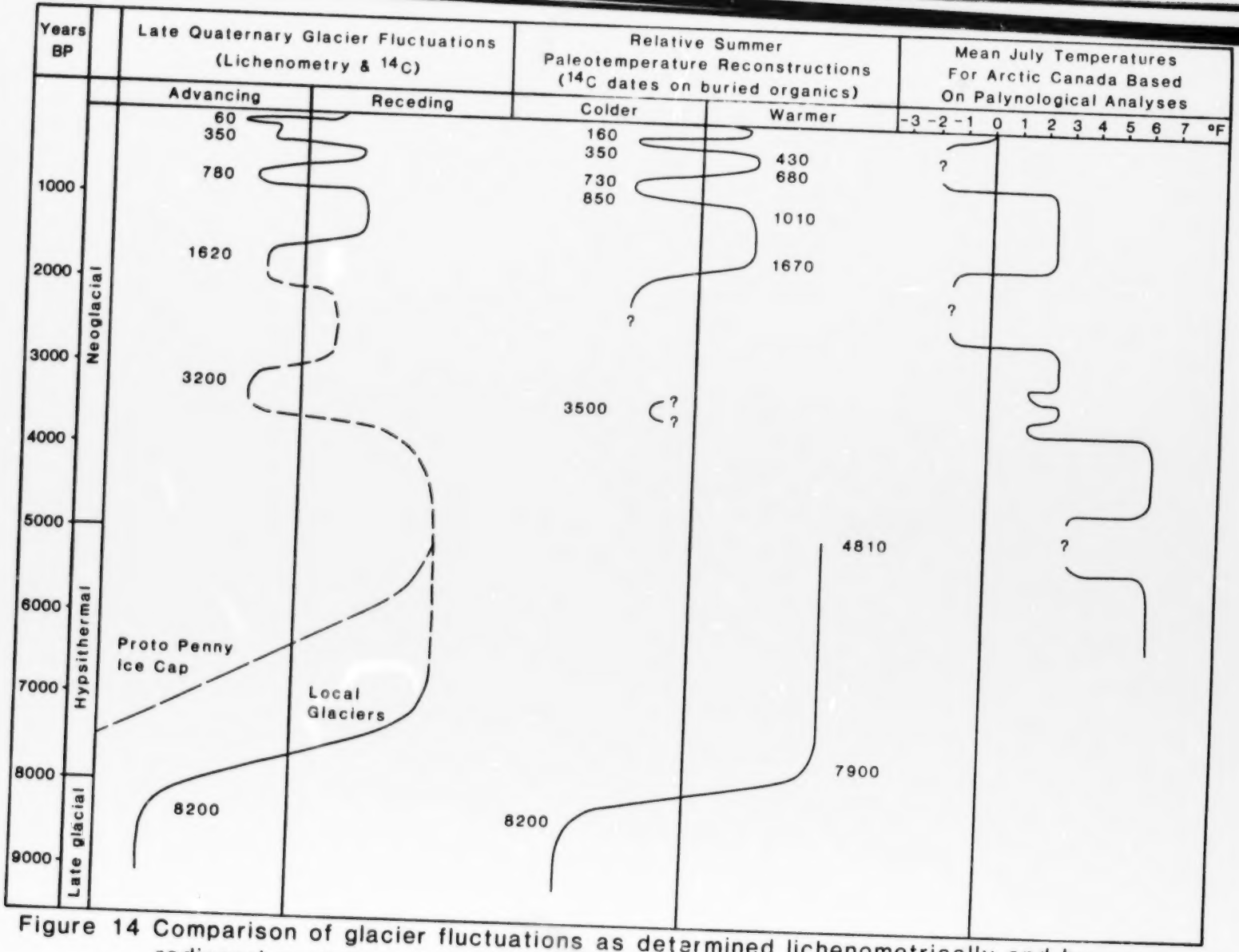


Figure 14 Comparison of glacier fluctuations as determined lichenometrically and by radiocarbon dating; summer paleotemperature reconstruction from radiocarbon dates on buried organic horizons; and mean July temperatures for Arctic Canada based on palynological analyses of Nichols, 1967; after Miller, 1973.

however, to date these studies remain localized in nature, and have not defined a consistent regional picture of the Quaternary history of the eastern Arctic. The study of the glacial processes and sediments continues today, both in onshore mapping surveys as well as in offshore bathymetric and hydrographic surveys.

5.3 Dating Methods

A wide variety of techniques has been used for developing a chronological sequence of glacial events for the eastern Arctic area. In general, these techniques provide either absolute or relative ages of glacial sediments or glacial features. These techniques can be used to determine relative age differences between either features or sediments, thereby defining the sequence of events which lead to their formation.

The methods or techniques which have been used in the eastern Arctic glacial studies can be grouped into three broad categories: 1) geological/geomorphic methods, 2) geochemical age dating, and 3) biological dating. The methods differ in that geological/geomorphic dating techniques and biological dating techniques do not provide absolute age determinations; these methods yield relative age estimates, although with 'calibration' by absolute dating methods, they can provide relatively accurate information on absolute age. The geochemical dating methods provide actual age determinations, based on changes in elemental isotope ratios. These methods provide a measure of the actual age in years before present (years BP).

The basic principle used in the geological/geomorphic dating techniques is the "law of superposition", which states that younger materials overlie older materials (assuming that sediments have not been deformed). If glacial erratics overlie unconsolidated deposits, it can be assumed that the deposits predate glaciation. A corollary to the law of superposition, is that surficial materials and glacial landforms decrease in age as one moves towards existing ice caps or active glaciers (Blake, 1976). Another geological dating method which defines relative age is based on the analysis of weathering zones (Table 4). In this case, weathering indices (such as pebble or cobble roundness, the relative proportions of pebbles or cobbles to boulders, or chemical weathering of rock surfaces) are used to indicate age differences or age groupings. The underlying assumption is that the greater the weathering of surficial materials, the older the surficial deposit. The abundance or grouping of foraminiferas (Feyling-Hanssen, 1976), or pollen grains (Short and Jacobs, 1982) also provides a means of correlating timed events in glacial deposits which are separated spatially. These techniques are particularly useful when calibrated by absolute dating methods - generally radiocarbon dating - and can provide detailed climatic and age records of limited sedimentary sequences. Detailed studies of mineralogical assemblages within the glacial sedimentary units can also provide an index of relative age. For example, analysis of etching on hornblende grains has proved to be a useful tool for determining relative ages of sediments, and if calibrated by an absolute dating technique, can be used to determine absolute ages of sediments (Table 4). Analysis of carbonate

Method	Materials Tested	Age Range of Technique	Selected References for Study Area
Radiocarbon Dating	Organic carbon (shells, wood)	200 to 40,000 BP	Blake, 1970; 1975, England, 1976a, Stuckenrath et al. 1979, Andrews, 1975,
Uranium Series Dating	Mollusk shells	2,000 to 500,000 BP	Kauffman et al. 1971,
Lichenometry	per cent lichen cover	0 to 8,000 BP	Miller, 1973, Koerner, 1980,
Oxygen Isotope Ratios	Changes in 18o/16o ratio in ice cores	0 to 20,000 BP	Fisher, 1979,
Amino Acid Ratios	Organics, forams, shells	0 to 400,000 BP	Andrews and Miller, 1976, Wehmiller and Hare, 1971,
Weathering Zones	Weathering character of surface rocks	$<$ 5,000 to $>$ 110,000 BP	Boyer and Pheasant, 1974, Ives, 1975,
Hornblende Grain Etching	Hornblende grains	5,000 to 200,000 BP	Locke, 1976.

Table 4 Methods used for establishing absolute ages of glacial events within the study area.

content in sediments has been used within the study area to determine directions of ice movement, from areas of high bedrock carbonate content to areas where the carbonate content is sparse (Andrews and Sim, 1964).

Geochemical age dating provides absolute ages of materials, and is a most useful tool in making age determinations. The most commonly used technique is that of radiocarbon dating (Table 4) where relative differences in elemental carbon isotopes provide a direct measure of age. The uranium series has also been used to provide limited dates from the study area, although these can be subject to large errors (Table 4, Kauffman et al., 1971). Other geochemical dating techniques do not provide absolute age estimates as such; however, enough correlations have been made to determine an adequate calibration curve. These techniques include determination of oxygen isotope ratios for ice cores, and determinations of amino acid ratios from organics in sediments (Table 4). Both techniques provide a high level of detail, and are very useful in making correlations between geological sequences which are spatially separated. The oxygen isotope method has been useful in providing information about relative temperatures at the time of ice or sediment deposition.

An interesting biological dating technique is that of lichenometry, which uses either the size or density of lichen growth to estimate age of surficial sediments (Table 4). The method assumes that lichen is destroyed by glacial ice cover and that it grows at a constant rate on surficial rocks and outcrops following exposure to the atmosphere. The method allows a high degree of resolution for recently exposed glacial deposits.

All of these methods have been used to determine either absolute or relative ages of glacial events and sediments in the eastern Arctic. It is important to note that only two absolute dating methods exist: radiocarbon dating and uranium series dating, and that other relative age dating methods are dependent to a large extent on being calibrated by these absolute methods. There are sources of errors with both absolute and relative dating methods, and at present, errors associated with many of the methods, such as weathering and amino acid ratios, are largely unknown. It must be emphasized that all methods are subject to some error, and that in some cases, this error may amount to a significant proportion of the absolute age of the material.

5.4 Pre-Quaternary Record

Information on pre-Quaternary events has been obtained from terrain not recently glaciated, and from buried organic deposits. It is probable that small glaciers and ice fields were present in Miocene Time (Late Tertiary) and continued into the pre-Wisconsin Period (Andrews and Miller, 1976). While little is known about the pre-Quaternary history of the eastern Arctic, other than that the large-scale physiography of Baffin Island probably remained unchanged during the Pleistocene, evidence does exist for both glacial and interglacial events (Craig and Fyles, 1960). The erosional

surface present in central and southern Baffin Island was probably tilted to the southwest during Tertiary time, with the resultant system of dendritic and parallel valleys accompanied by a westward migration of the drainage divide. It appears that the last pre-Wisconsin ice sheet supported active glaciers in the main fiords of the northeastern coast of Baffin Island, and formed piedmont glaciers on the coastal lowlands. Glaciers with deep outlet valleys may have contributed to the migration of the drainage divide and dissected the eastern mountain rim.

5.5 Quaternary Record

Andrews et al. (1975) suggest that the stades that occurred earlier than 40,000 years BP were localized in nature, and that the overall extent of the Laurentide Ice Sheet did not alter significantly during the period 100,000 to 40,000 years BP. They note, however, that between 40,000 and 8,000 years BP, there was a considerable reduction in the volume of the northeastern sector of the Laurentide Ice Sheet (Andrews and Ives, 1978).

The glacial record for the eastern and southeastern Queen Elizabeth Islands has been obscured primarily by the advance of Laurentide ice over the area during the Wisconsin glaciation. England (1976a) suggests that the last major glaciation over this area was associated with a series of convergent ice fields that formed an ice complex, rather than an ice sheet. It is generally agreed that the major channels that presently exist in the southeastern Queen Elizabeth Islands were ice-free by 8,000 years BP (Blake, 1976). However, fiord heads in many areas did not become free of ice until between 6,300 and 5,000 years BP (Andrews and Ives, 1978).

The glaciation of northern Baffin Island is not fully documented at this time. Clearly, ice from the west and southwest overrode parts of Bylot Island (Klassen, 1981), and it appears that Lancaster Sound was a major conduit for the movement of ice from the west towards the east into Baffin Bay. Bird (1980) suggests that Laurentide ice did not cross Parry Channel (Lancaster Sound) into the Queen Elizabeth Islands.

During the last glacial maximum, it is possible that ice from local caps on Brodeur and Borden peninsulas, valley glaciers flowing northward along Admiralty and Navy Board inlets, and possibly ice from Foxe Basin, merged with ice from the Devon ice cap in Lancaster Sound. Pond and Navy Board inlets served as major outlets for ice moving across, or originating on, northern Baffin Island (Klassen, pers. com., 1980). Segments of lateral moraines between Mount Herodier and Cape Macculloch on northern Baffin Island mark ice-marginal positions during the one or more glaciations which filled Pond Inlet with ice. On northeastern Baffin Island, the arrangement of lateral moraines suggests that ice moving through fiords overrode much of the foreland and inner continental shelf, and in the latter case, either by spreading out as lobes or by converging and dissecting existing troughs (Hodgson and Haselton, 1974).

On Bylot Island, most glacial evidence indicates a general expansion of existing intermontane glaciers. However, a moraine ridge swinging inland from Cape Byam Martin at Bathurst Bay, with an ice contact delta at its inland margin, may denote a stage of a glacier flowing from Lancaster Sound into Baffin Bay. Also, an arcuate flight of moraines located at the northwestern tip of the island may mark the terminus of ice flowing from Baffin Island and Eclipse Sound during a later stage of ice retreat. A series of high-level moraines on the northeastern coast of Bylot Island marks the contact of intermontane ice with Lancaster Sound ice, and may indicate more than one advance from Lancaster Sound. Klassen (1982) suggests that on Bylot Island, native glaciers have recently (less than 120 years BP) advanced, although most are retreating. At their recent maximum, they are considered to have been as far advanced as at any time since before 39,000 years BP.

The first major event of the Wisconsin glaciation on Baffin Island is a readvance that has been generally dated between 80,000 and 115,000 years BP. This advance is referred to as the Clyde River Stade or the Alikdjuak Stade, and lasted for approximately 10,000 years (Table 3). Moraines and marine or glaciomarine deltas formed at this time are generally in the order of 80 m above present sea level (Pheasant and Andrews, 1973). This advance was followed by an interglacial period, but the record subsequently becomes less clear, and correlations between areas have not been made definitively for the remainder of the Wisconsin.

In the Clyde River area, the Ayr Lake readvance is considered to be significantly older than the Napiat advance that characterizes northern Cumberland Peninsula. At the time of the Ayr Lake glaciation, the ice front would have been seaward of the coast and would have consisted of tidewater glaciers of ice sheets (Fillon et al., 1981). The subsequent period, between about 50,000 and 10,000 years BP, is considered to be one of numerous local glaciers that formed moraines, and this period is referred to by Andrews and Barry (1978) as an interstadial complex.

Fillon et al. (1981) suggest that each of the ice advances, subsequent to the Clyde River Stade, were successively less extensive (Figure 15). However, these would have extended to lower elevations, in terms of topography, due to the relative lowering of sea levels throughout the Wisconsin. Miller et al. (1977) suggest that warm waters invaded the Baffin area during the Quajon Interstadial period. The sequence of events since approximately 50,000 years BP is poorly known, and a key element that is lacking in the interpretation of the middle and late Wisconsin glacial events is the presence of glacial deposits such as moraines.

Following the late Wisconsin Interstade, a readvance took place bewtween 10,000 and 8,000 years BP. This resulted in the development of an extensive (Cockburn) moraine system (Figure 16) that stretches from southeastern to northwestern Baffin Island along the eastern coast. The Cockburn moraine system has been redefined by Andrews and Ives (1978) as the Baffinland Drift, and was formed during the Cockburn substage of the Baffinland Stade.

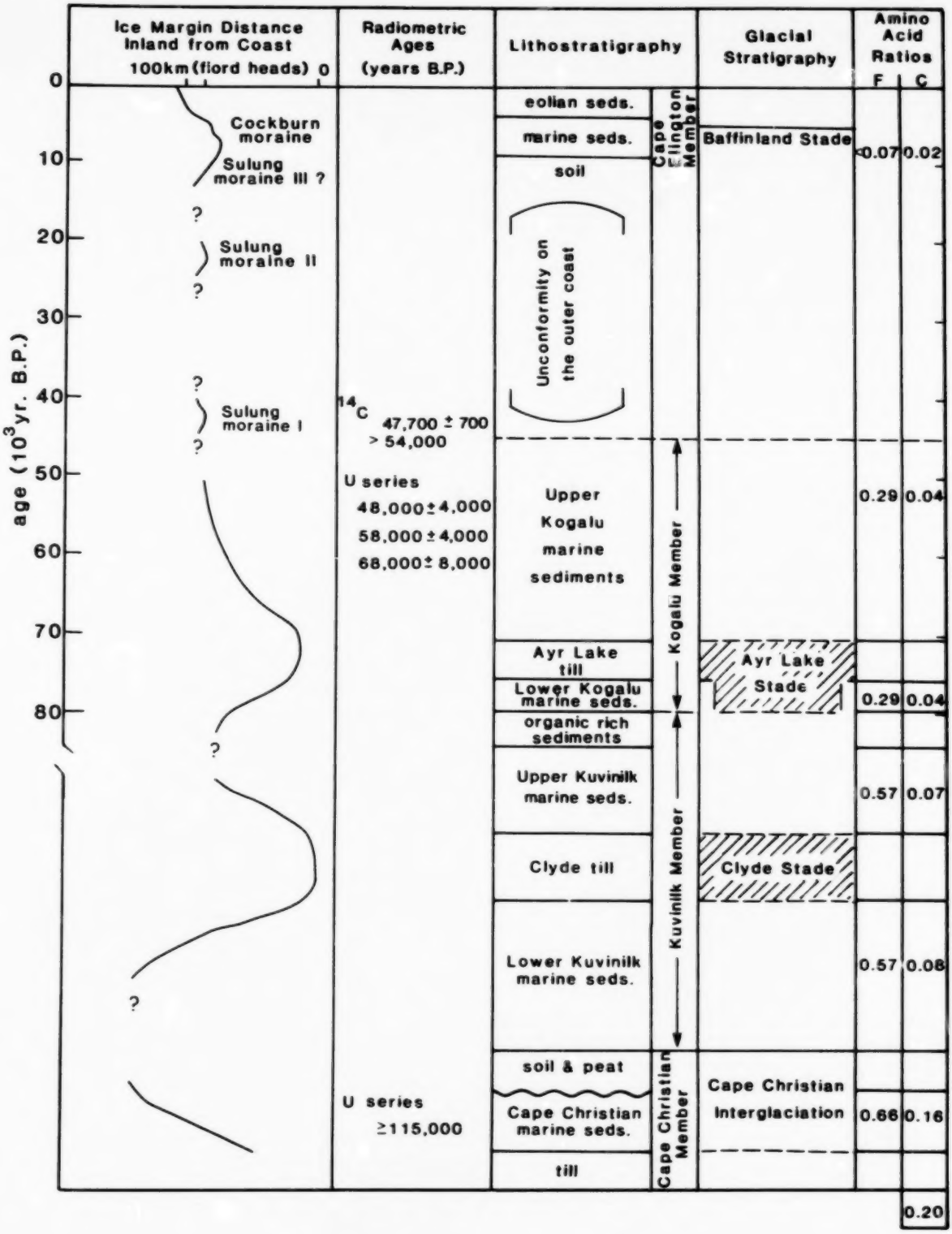


Figure 15 Time distance diagram for ice on eastern Baffin Island, after Fillon et al. 1981

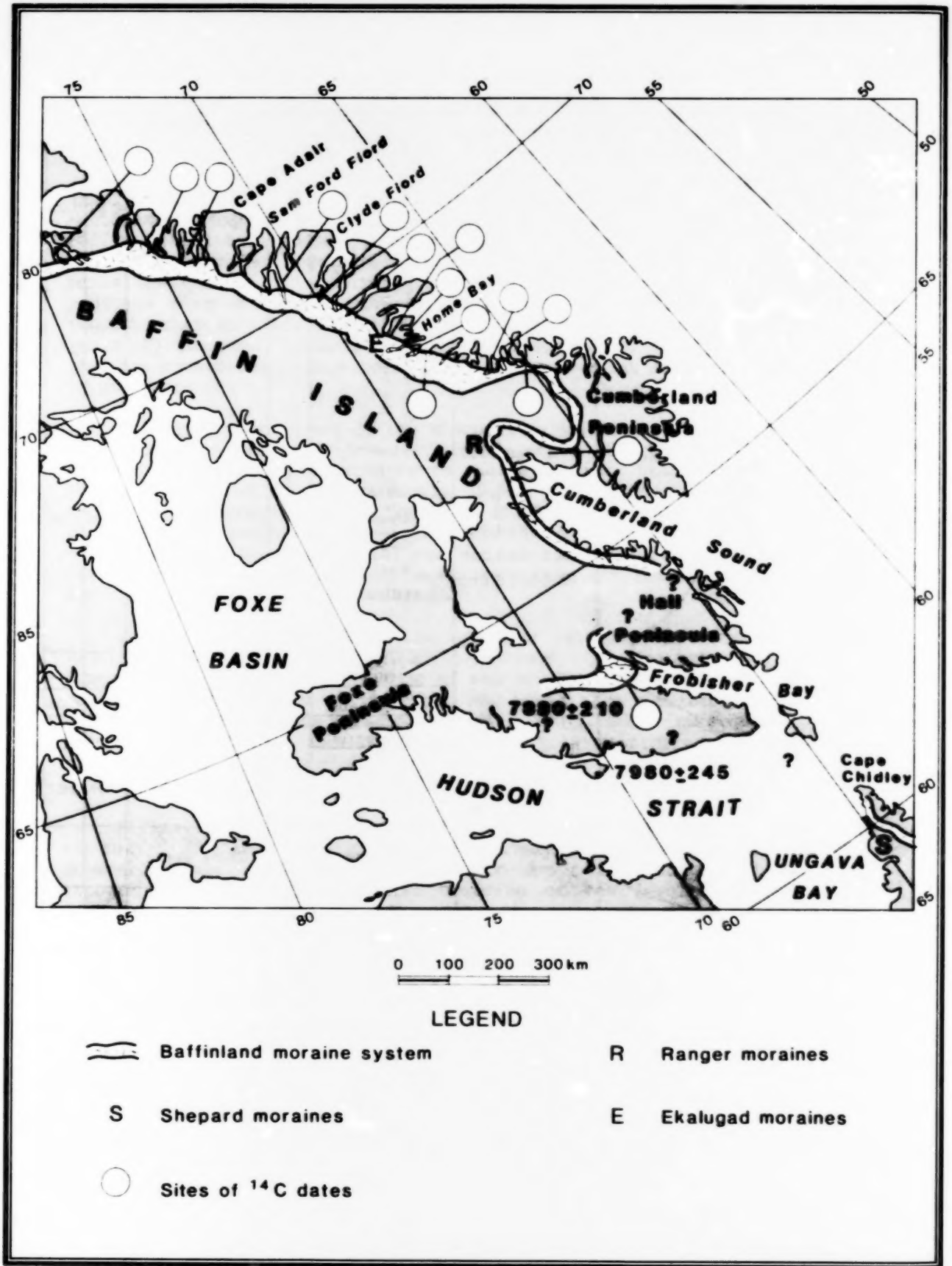


Fig. 16 Location of the Cockburn end moraine system and radiocarbon dates associated with the moraines, after Andrews and Ives, 1978.

Before this readvance, there had been a major interstadial period that involved a significant reduction of ice area and volume, such that the readvance lay well inland of the outer coast (Fillon et al., 1981). Numerous neoglacial events have been recorded since the formation of Cockburn moraines. Essentially, a hypsithermal warm period between 8,000 and 4,000 years BP was followed by a series of readvances (Miller, 1973; Locke, 1976). A significant characteristic of the most recent advances, that is during the past 1,000 years, is that these have been as extensive as those of the last glacial maximum, before 40,000 years BP (Miller, 1976). Miller (1973) describes a sequence that involves retreat of the late Wisconsin ice sheet from fiords and cirques such that by 5,500 years BP, the fiord glaciers had melted behind the present Penny Ice Cap. Following the retreat of major outlet glaciers, local snow fields pushed glacier tongues into the fiords to breach the lateral moraines that had been left by former valley glaciers. Church (1972) suggests that sandurs began to develop after 7,000 years BP at the fiord heads during the warm interval. Church also notes that between 1950 and 1970, the ice retreated approximately five km on Baffin Island. The present glaciated area is in the order of 37,000 km^2, versus 140,000 km^2 approximately 300 years ago (Hattersly-Smith, 1974). Blake (1978) suggests that at the present time, the outlet glaciers of southeastern Ellesmere Island are undergoing retreat following a slight advance.

5.6 Sea Level Changes

The relative sea level changes have been documented in great detail and with a relatively high level of accuracy on the southern coast of Ellesmere Island (Blake, 1975). As with other areas of the eastern Queen Elizabeth Islands, data before 10,000 years BP are lacking, primarily due to the effects of extensive late Wisconsin glaciation. Blake indicates that emergence has been relatively continuous since 9,500 years BP, and that there have been no oscillations of sea level greater than two m with durations longer than 500 years during that period. The only major interruption in the emergence that continued throughout this period is a possible relative submergence at approximately 5,000 years BP.

Initial emergence (9,000 to 8,000 years BP) was very rapid in southern Ellesmere, in the order of 7.0 m/century, and marine overlap at this time was 130 m above the present sea level at Cape Storm. The rate of emergence slowed to 0.8 m/century between 6,500 and 4,500 years BP, and from 2,400 years BP to the present time, the rate of relative sea level change has been less than 0.3 m/century. Other data from the eastern Queen Elizabeth Islands, reviewed by England (1976b) show that the northeastern Devon Island coast has emerged approximately 40 m since 7,500 years BP, and 20 m since 6,000 years BP. The emergence curve for eastern Devon Island, therefore, apparently shows the same trend as those for southern Ellesmere Island, with a rapid emergence rate that decreases between 5,000 and 4,000 years BP. The pattern of sea level changes since the Wisconsin is generally poorly documented throughout eastern Baffin Island. There is some general

agreement that relative sea level was at approximately +80 m between 100,000 and 70,000 years BP. There is also general agreement that since 8,000 years BP, there has been a steady glacio-isostatic rebound that has resulted in the steady rise of the land with respect to the sea. Andrews (1979) suggests that immediately prior to the Cockburn Stade, there were several submergence periods, although these have not been documented properly. Andrews (1979) further suggests that there has been a marine submergence on the eastern Baffin Island coast over the past 1,000 years.

5.7 Relationship of Wisconsin Glacial Events in the Eastern Arctic to Events in Southern Canada

The sequence of events which led to the glaciations of the eastern Arctic are in a broad sense related to the larger scale events of the continental glaciation in southern Canada and northern United States. It should be emphasized that the study area has always been located near the margin of the ice sheet, and as such, the sediments record numerous small scale fluctuations of the ice margin that are not well documented in southern areas because most of the land surface area was beneath the ice. In many areas on eastern Baffin Island, the distance between the Wisconsin glacial maximum and neoglacial events can be measured in tens of kilometres (Andrews and Miller, 1976).

Ives and Andrews (1963) hypothesized the following sequence of events for the Wisconsin glaciation in the eastern Arctic:

1) Initial development of an ice sheet on the eastern flank of central Baffin Island.

2) Expansion of the ice sheet westward into Foxe Basin with a subsequent development of a dispersal centre within Foxe Basin.

3) The Foxe Basin dispersal centre is thought to have remained active during the Wisconsin, at some point joining with the continental ice sheets to the south and west (Prest, 1976).

4) At an unknown point in time, the Foxe Basin dispersal centre increased in size enough to cause a reversal of ice flow from west to east across central Baffin Island (Ives and Andrews, 1963). This flow trend is suggested by the distribution of carbonate which presumably originates from limestone outcrops in Foxe Basin (Andrews and Sim, 1964). It is possible that the westward flow across Baffin Island coincided with the maximum glacial extent on eastern Baffin Island, the Clyde Stade, during early Wisconsin (Figures 16 and 17).

5) At some point during early to middle Wisconsin, it appears that the Baffin Island ice sheet and the main Laurentide ice sheet were uncoupled. This relationship is suggested by the fact that the eastern Baffin Island ice margin position continually retreated landward during middle to late

Figure 17 Maximum extent of the Laurentide ice sheet along eastern Baffin Island, after Ives, 1978.

Wisconsin and decreased substantially in ice volume (Fillon et al., 1981). At the same time, the main Laurentide ice sheet increased continuously in volume with a late Wisconsin maximum at approximately 17,000 to 15,000 years BP (Prest, 1976).

6) The late Wisconsin Cockburn Stade (Figure 16), which took place on Baffin Island around 8,500 years BP appears to be a result of expansion of an ice cap centred on Baffin Island, creating flow both to the east and west. This interstade may have an analog in the Laurentide ice movement around 8,000 years BP when minor readvance occurred.

7) The relationship of these glacial events to the Innuitian Ice Complex is uncertain because of the controversy surrounding both the extent of the Innuitian Ice Complex and the ages of Wisconsin maxima in the northern area.

8) Small-scale neoglaciation events have characterized the last 8,000 years of deglaciation, with the recent century being dominated by wide-scale glacial retreat to present ice margin positions. It is uncertain if these neoglacial events from the eastern Arctic are also represented by minor fluctuations of the Laurentide Ice Sheet as it retreated.

9) Fillon et al., (1981) speculate that variations in offshore sedimentation patterns may be related to meso-scale circulation changes in the Labrador Sea during the Wisconsin, which controlled offshore water temperatures and iceberg drift patterns.

5.8 Man

Ancestors of the Inuit began to occupy the Arctic Archipelago during the Recent or Holocene Epoch. The Denbigh people, the first Eskimo culture of the Canadian Arctic, spread rapidly eastward from Alaska about 5,000 years ago and were known to exist on Ellesmere Island 4,500 years ago. It is thought that a cooling climatic trend forced the Denbigh people south.

When warmer conditions returned approximately 3,000 years ago, the Dorset Culture, or Tconit, moved across the Arctic and reoccupied most of the region originally inhabited by the Denbigh people. The Tconit reached a peak about 2,600 years ago. Their ancestors probably migrated from Alaska. The Dorset or Tconit people are known for their conspicuously small tools. At their peak, they occupied a surprisingly vast area, from Banks Island in the west, to Greenland in the east, and from Ellesmere Island in the north, to Newfoundland in the south. They became extinct, for unknown reasons, about 600 years ago.

Approximately 1,000 to 1,500 years ago, a second warm climatic trend coincided with the eastward expansion of the Thule culture in the Arctic. This group developed from the original Denbigh people in Alaska. After centuries during which the Dorset people mingled with, or retreated from the Thule people, the Dorset culture disappeared, and from the Thule culture developed the present day Inuit.

6 The Seismicity of the Eastern Arctic

Seismicity is the phenomenon of earth movements. The earthquake history of northern Canada is brief and largely incomplete before 1962. Before 1950, Arctic seismicity was monitored almost exclusively through a world network of seismographs established during the early 1900's. Only major arctic earthquakes with magnitude 6 or greater could be detected and located by the world network, because of the small number of stations and low instrument resolution (Stevens, 1974).

6.1 The Arctic Record

Improved monitoring of eastern Arctic seismicity took place after 1950 when the first Arctic seismograph station opened at Resolute. Equipped with short period vertical seismographs, this station was capable of recording earthquakes of magnitude 6 or greater within epicentral distances of 2,000 km. Even after the opening of this station, few epicentres could be located accurately because of the large distances and low magnifications of other stations operating in southern Canada, Alaska, or Greenland. The inability to resolve the sources of the numerous local seismic events recorded at Resolute caused uncertainties as to whether the events represented real earth tremors, or were the result of nearby ice activity (Milne and Smith, 1961).

Leblanc and Wetmiller (1974) and Stevens (1974) conclude, after a review of data for 1899 to 1961, that earthquakes recorded during the early part of this time period can have epicentral uncertainties of up to 300 km. After 1930, the resolution was improved to about 100 km with the expansion of the world seismograph network and development of instruments with lower threshold detection (Basham et al., 1977). After 1961, establishment of seismograph stations at Alert and Mould Bay confirmed that the pattern of local seismicity was due to earthquakes and not to ice movements. Between 1962 and 1971, five other stations were added to the northern Canadian network, with the result that most points within the Arctic islands are within 800 km from a recording system. The eastern Arctic, particularly Baffin Island, is not as well monitored, as it is more distant from all recording stations.

Now, every station north of 60°N has instruments capable of recording events of at least magnitude 4 (Stevens and Milne, 1973), although epicentres are often difficult to locate accurately because of the position of the various stations.

Since 1962, significant improvements in the sensitivity of the world seismograph network, the computer processing of seismic data, and the expansion of the northern Canadian seismograph network have resulted in much better detection of arctic earthquakes. There are more than 3,400 Arctic earthquakes known for the period 1908-1975, of which more than 80 per cent have been located since 1962 (Wetmiller and Forsyth, 1978).

The accuracy of arctic earthquake data remains variable from continental to the deep oceanic areas. For the continental areas, Wetmiller and Forsyth (1978) suggest that all earthquakes of magnitude 4.5 and greater have been detected since 1960, and all of magnitude 3.7 or greater since 1968. Within the deeper oceanic areas, 4.5 is set as the threshold magnitude for complete detection. Coverage of oceanic events of lower magnitudes is only reliable along continental margins close to seismograph stations. Basham et al. (1977) conclude that larger continental earthquakes detected by several stations have epicentral uncertainties of about 50 km, while smaller magnitude events could be mislocated by as much as 100 km. The present epicentral reliability of earthquakes that originate in the deeper oceanic regions is difficult to estimate, but is probably in the range of 50 km.

The low detection threshold reached in 1962 has greatly increased the earthquake data file. These data confirm the seismic nature of some areas and delineated additional clusters and trends not previously apparent. However these trends represent only 14 years of data, which is too short in geological terms to define clearly the relationship of seismicity to current tectonic forces. The differences in threshold detection levels between continental and oceanic areas can also lead to errors in interpretations. While many smaller earthquakes are recorded in continental areas, the presence or absence of similar events in the deeper ocean areas is unknown (Wetmiller and Forsyth, 1978). It should be remembered that the level of monitoring of certain areas of the eastern Arctic, particularly northern Baffin Island, is not as high as the rest of the Arctic islands because of the distribution of the recording stations.

6.2 Distribution of Earthquake Epicentres and Magnitudes of Earthquakes

The locations of epicentres recorded in the Canadian Arctic from 1899 to 1961 (Figure 18) indicate that the eastern Arctic is seismically active (Stevens, 1974; and Basham et al., 1977). Although individual epicentres may be associated with large uncertainties, there is a concentration of earthquakes in the Arctic.

The improvements in stations and detecting capabilities implemented in 1962 enabled patterns of seismicity to be delineated for the eastern Arctic (Figure 19). The data indicate areas of activity near the larger earthquakes located teleseismically before 1962, and additional clusters and trends not previously apparent (Wetmiller, 1974).

Calculations used in determining the distribution of seismic risk in portions of eastern and western Canada by Milne and Davenport (1969), formed the basis of the 1970 "National Building Code Seismic Zonation Map of Canada". The four seismic zones are described by different levels of horizontal ground accelerations, expressed as a percentage of gravitational acceleration (per centg), that may be exceeded each year with a probability of one in 100. The eastern Arctic falls largely into the area of highest risk, where minor earthquakes are frequent, and potentially damaging earthquakes may occur several times in a 100 year period (Stevens, 1974; and Whitam, 1975).

Figure 18 Earthquake epicentres in northern Canada and adjacent regions (north of 59°N) for the time period 1899-1961, after Basham et al.1977.

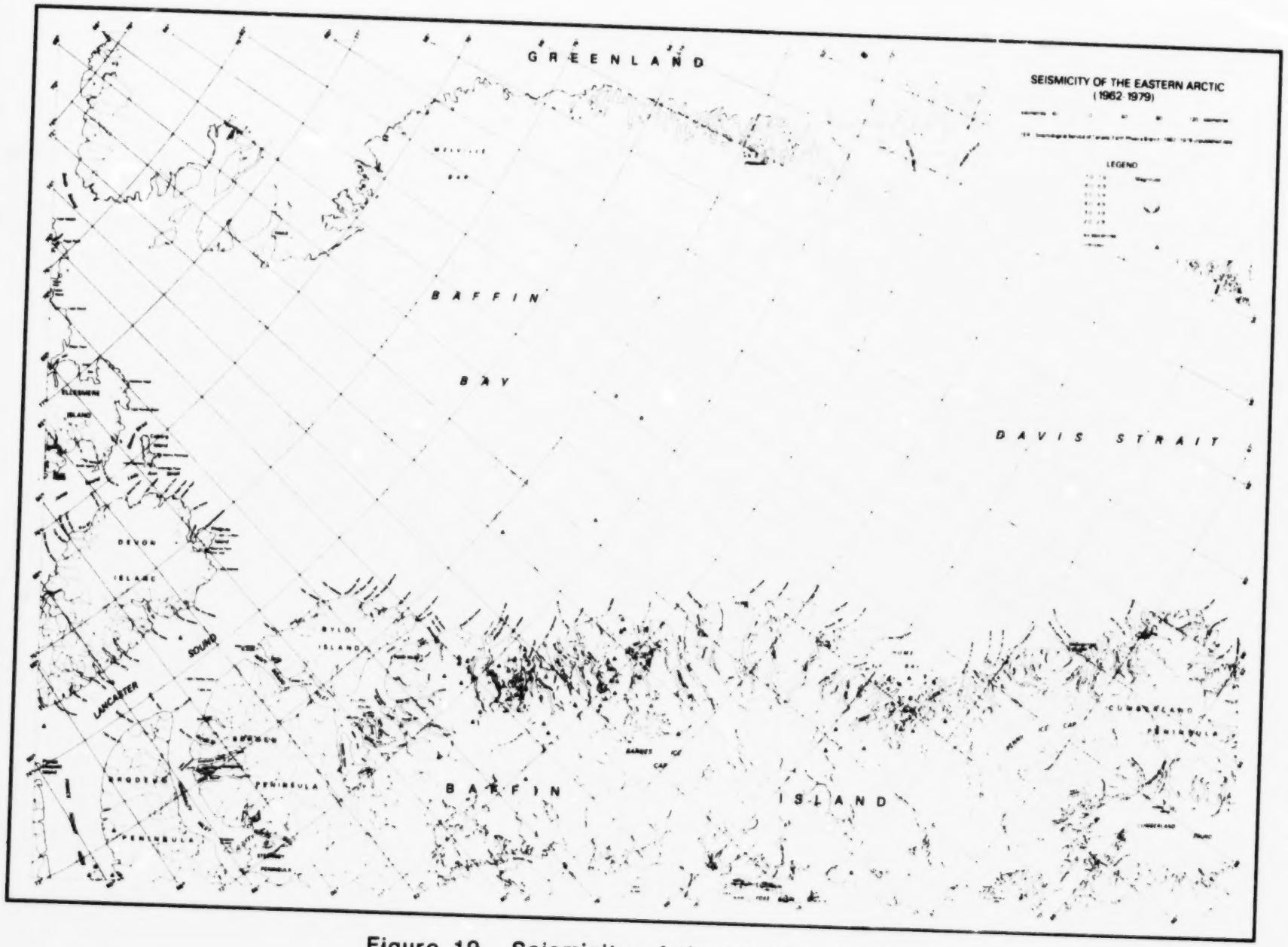


Figure 19 Seismicity of the eastern Arctic.

Seismic activity in the eastern Arctic is concentrated in northwestern Baffin Bay, along the southern coast of Devon Island, and along the eastern coast of Baffin Island (Figure 19). More scattered activity occurs throughout Lancaster Sound, western central Baffin Bay, and southern Baffin Island. Pre-1965 earthquakes of magnitude 5 and greater show an apparent trend along the west central portion of Baffin Bay. However, this trend is based on very limited data and is not apparent in more recent earthquake activity (Basham et al., 1977). The main part of low level seismic activity since 1962 in Baffin Bay has occurred on the continental slopes of Baffin Island and Greenland, largely within the 2,000 m bathymetric contour. On Baffin Island, earthquake activity is concentrated in two distinct areas, near Buchan Gulf and Home Bay, extending from the offshore regions to the centre of the island. The area to the southeast of Buchan Gulf was the site of a series of earthquakes in 1963, and the area to the northwest in 1969.

Davis and Nares straits appear to be essentially aseismic. Although Davis Strait is far from any station, Nares Strait is well monitored at Alert. While earthquakes occur regularly each year in the eastern Arctic, the events recorded are usually of magnitude 5 or lower. However, earthquakes of larger magnitudes have occurred. In 1933, north central Baffin Bay was the site of the largest known earthquake in the Canadian Arctic (magnitude 7.3). Since then, earthquakes of magnitudes 6 to 6.5 have occurred in 1934, 1945, 1947, 1957 and 1963 (Basham et al., 1977). Earthquakes of magnitudes 6 or greater are considered large and potentially destructive (Stevens and Milne, 1973).

6.3 Earthquake Mechanisms

The eastern Arctic lies in an area that is, in general, poorly understood in terms of global tectonics, and as a result, the forces responsible for the current low level of seismic activity are largely speculative. Seismic activity may be the result of stresses produced by plate movements, glacial unloading, uncompensated sedimentary loads, or diapiric structures.

The only direct evidence of the stress field in the eastern Arctic resulted from studies by Hashizune (1973) on mechanisms for two earthquakes that occurred under Baffin Island in 1970 and 1972. These earthquakes were associated with normal and slip-strike faults, and the analysis showed that the tectonic stress may have been tensional in a northeast-southwest direction, normal to the general trend of Baffin Island. The tectonic implication of these limited observations is unclear. These earthquakes may have been produced by a regional stress field such as the one described above, or by local stress concentrations associated with geologic structures (Basham et al., 1977).

Hashizune (1973) and Herron et al. (1974) suggest that Baffin Bay, Nares Strait, and Davis Strait were created by a separation of Baffin Island and Greenland during the Cenozoic Era (maximum 65 million years ago).

However, geophysical investigations by Wetmiller (1973) and Grant (1975) demonstrated the presence of a bridge of continental crust across northern Baffin Bay and the possibility of continuous continental crust beneath the Davis Strait sill. Kerr (1980) suggests that there was a small movement (25 km) along Nares Strait combined with rotation, which stretched, thinned and foundered a large area of continental crust. This theory is in general agreement with the fact that contemporary seismicity does not occur in a continuous zone from the Labrador Sea through the Baffin region to the Arctic Ocean as would be expected if an ocean ridge was present. Instead, it appears to be concentrated in northern Baffin Bay, along the northeastern coast of Baffin Island, and in the Labrador Sea, with distinct gaps in Nares and Davis straits. Also, the absence of large subcrustal earthquakes suggests that tectonic forces characteristic of plate margins are not acting directly in the eastern Arctic.

The areas of seismicity adjacent to Home Bay and Buchan Gulf on Baffin Island are areas where structures may have been reactivated by rifting or foundering processes. The present seismicity correlates well with the presence of diabase dikes. These intrusions are not necessarily the sources of the present seismicity, but deeper structures are perhaps being reactivated by contemporary stresses. Based on the earthquakes studied by Hashizune (1973), these structures may be reactivated at relatively shallow depths (five to ten km) (Basham et al., 1977).

None of the preceding theories adequately explains the seismicity of the eastern Arctic. Epicentral trends present in the eastern Arctic also correlate well with areas of intense isostatic readjustments (Walcott, 1970), and coincides with areas where maximum stress gradients may exist. The tilting to the northeast of the Baffin Island-Foxe Basin is such that the northeastern coast of Baffin Island is experiencing a higher rate of uplifting than the area to the southwest, and may be acting as a hinge zone (Lambert and Vanicek, 1979; and Basham et al., 1977). It is possible that if these areas are isostatically unbalanced, stresses due to glacial unloading may be triggering low-level seismic activity on pre-existing structures.

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